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Assessment of the rainfall response on headwater streams in lower coastal plain South Carolina

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ASSESSMENT OF THE RAINFALL RESPONSE ON HEADWATER STREAMS IN LOWER COASTAL PLAIN SOUTH CAROLINA

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ABSTRACT

Toward the goal of the assessment of the rainfall response on forested headwater catchments in the Lower Coastal Plain (LCP) of South Carolina, rainfall, streamflow, and groundwater elevation were monitored on two similar streams. Upper Debidue Creek (UDC) in coastal Georgetown County, South Carolina is slated for development and Watershed 80 (WS80) in the Francis Marion National Forest serves as an undeveloped reference watershed. Spatial rainfall variability was assessed at UDC and it was concluded that a single gage was sufficient to accurately measure rainfall for this watershed. Throughfall measurements at UDC indicate a seasonal difference that may influence seasonal trends in streamflow. The rainfall response on the two watersheds was measured as total storm flow and direct runoff components of watershed outflows. Storm event runoff was determined by a graphical hydrograph separation method that takes into account the unique mechanisms of runoff generation in the LCP related to low topography and shallow water table. Variability in runoff generation at UDC and WS80 was related to seasonal trends of evapotranspiration that determine soil moisture conditions and are related to seasonal fluctuations in groundwater elevation. Break point water table elevations were determined for each watershed above which runoff generation was observed to increase sharply. The SCS Curve Number method for runoff modeling was compared to measured rainfall and runoff for storm events on both watersheds. Parameter selection by the accepted methodology does not appear to accurately model runoff generation on these LCP headwater catchments. The strong relationship between groundwater elevation and runoff generation should be considered

for applications of the Curve Number method in similar watersheds. The effect of seasonal trends in groundwater elevation on the rainfall response for similar streams in the LCP may not be well modeled by the median measure for runoff generation that is typically used due to fluctuating moisture conditions.

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CHAPTER ONE

INTRODUCTION

The population has been increasing in coastal South Carolina and this trend is expected to continue and contribute to large increases in development in the lower coastal plain (LCP) region of the state over the next 30 years (Allen and Lu, 2003). While development is good for the economic interests of an area, it is often accompanied by land cover changes that have adverse effects on local watersheds. Impervious cover typically increases when land development takes place, and the percentage of impervious cover in a watershed is now used as an environmental indicator to assist local authorities with resource protection planning (Arnold and Gibbons, 1996). Land cover change associated with land development changes runoff patterns and alters local streamflows and this can result in declining biological health in developing watersheds (Booth et al., 2002). Water quality often declines in developing watersheds as well due to increasing non-point source runoff and this has been demonstrated in the sensitive tidal marsh ecosystems in the LCP (Holland et al., 2004). As development increases and impervious cover does as well storm runoff will increase in volume and peak discharge, altering coastal pre-development site hydrology and creating issues downstream (Blair et al., 2011). In order to accommodate increasing development in the LCP, adequate stormwater management measures must accompany land cover changes to ensure the continued health of area watersheds. The performance of such stormwater handling measures depends on data-driven design that accounts for the expected rainfall response for area watersheds based on pre-development runoff generation. This situation is complicated in the LCP because of unique hydrologic conditions that differ from the better studied hydrology of upland watersheds.

The LCP of South Carolina has unique hydrologic conditions that differ from higher gradient watersheds. The area is characterized by very flat topography and a shallow water table. The water table position changes in response to seasonal trends in evapotranspiration (ET). This contributes to variable moisture levels that determine outflow production and the response to rainfall (Amatya et al., 2006; Harder et al., 2007; La Torre Torres et al., 2011; Sun et al., 2002; Williams, 2007). High ET during the summer months lowers the water table, increasing available soil storage. Runoff generation is lower during these months, and watershed outflows become intermittent on headwater streams when the water table is lowered sufficiently. Declining ET during the fall months results in replenishment of the groundwater aquifer that raises the water table position when rainfall occurs. During the winter months when ET is lowest, the water table position remains high. This contributes to higher watershed outflows and higher runoff generation during these months. Sustained baseflows on these headwater streams are the result of groundwater contributions. Headwater streams in the LCP function as natural gravity-driven drainage for the groundwater aquifer (Amatya et al., 2006). Sustained outflows not related directly to the rainfall response are a function of the water table position. Therefore, these outflows vary on an annual basis accordingly.

Runoff generation in the rainfall response on headwater catchments in the LCP ranges according to these variable conditions as well. This process is depicted in Figure 1.1 for the dormant season and Figure 1.2 for the growing season.

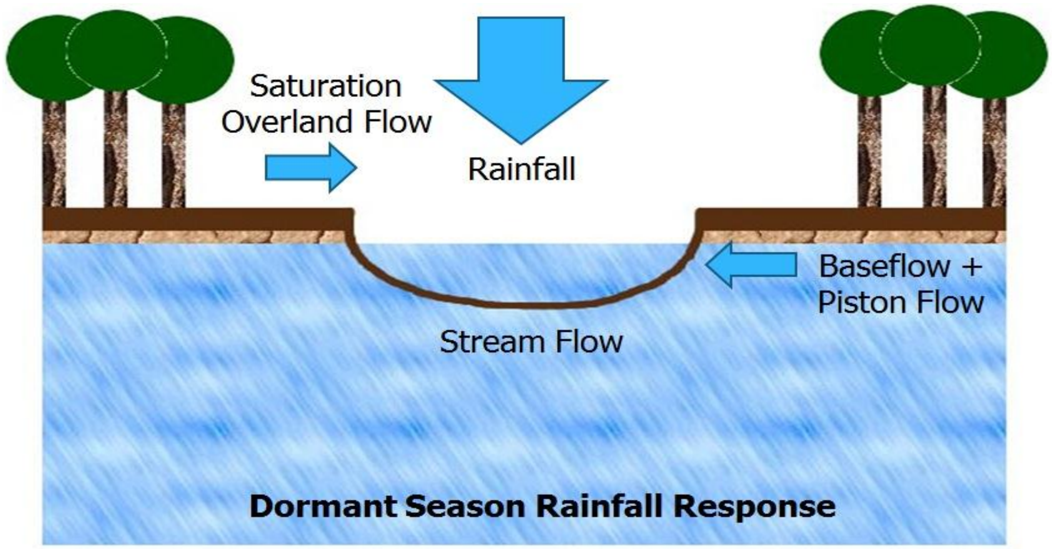

Figure 1.1. Dormant season rainfall response and runoff generation typical of the Lower Coastal Plain of South Carolina.

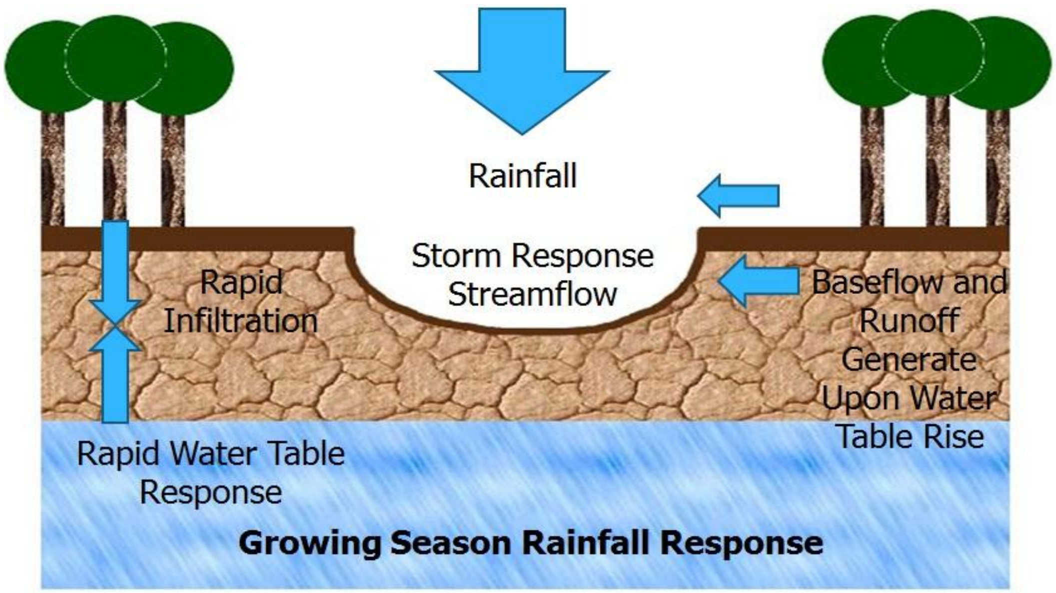

Figure 1.2. Growing season rainfall response and runoff generation typical of the Lower Coastal Plain of South Carolina.

Annual outflow depths measured as a percentage of annual precipitation range from less than 10% to greater than 50% for LCP headwater streams (Amatya et al., 2006; Harder et al., 2007). Runoff generation is not defined by rainfall alone. Antecedent moisture conditions (AMC) that are determined by soil moisture characteristics prior to rainfall play a large role in determining runoff for any given storm event (Amatya et al., 2006; Harder et al., 2007; La Torre Torres et al., 2011). Seasonal trends in ET contribute to trends in AMC, with wet AMC persisting during winter months and dry AMC during summer months. Runoff generation mechanisms in the LCP differ from upland catchments where higher gradients determine outflows (Sun et al., 2002). Saturation excess overland flow is the dominant mechanism in LCP forested headwater catchments due to high infiltration rates and low gradient conditions (Williams, 2007; La Torre Torres et al., 2011). This process is influenced by water table elevations which have been shown to respond rapidly to rainfall (Williams, 1978). Saturated areas that generate runoff are determined by groundwater elevation and these areas vary in size between storms according to AMC. These variable source areas contribute to differences in runoff generation between storm events in lowland watersheds (Hewlett and Hibbert, 1965; Eshleman et al., 1994). The relationship between groundwater elevation and surface water generation in response to rainfall may also contribute to accelerated baseflow contributions to outflow. Rapid water table rise at the watershed divide once rainfall begins is thought to increase subsurface groundwater discharges towards the stream (Williams, 2007). Groundwater discharges toward the stream can also be increased by the piston-flow mechanism when the capillary rise of groundwater extends above the ground surface elevation and rainfall transmits pressure through the aquifer towards the stream (Williams, 2007). Shallow water table hydrology and low topography in the LCP present challenges because it is difficult to directly measure the groundwater contributions to outflows and the rainfall response varies for any given storm event.

OBJECTIVES

In order to provide a better understanding of baseline hydrology and the rainfall response for headwater catchments in the LCP of South Carolina, this work aims to

1. Assess spatial rainfall variability and canopy interception in a forested headwater catchment to refine water budgets and determine the role that canopy cover has on seasonal moisture trends,

2. Measure the rainfall response on two comparable LCP headwater catchments and assess the relationship between runoff generation and seasonal trends in antecedent moisture,

3. And compare runoff estimates modeled by the SCS Curve Number method to storm event data on two LCP headwater catchments to assess parameter selection by accepted methodology.

Results of this work will contribute to a better understanding of runoff generation and the relationship that seasonal trends in moisture conditions have on streamflows for LCP headwater streams. This will provide better guidance for stormwater handling and management as development and land cover change continues to increase in the LCP.

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CHAPTER TWO

Literature Review

The population has been increasing in coastal South Carolina and this trend is expected to continue and contribute to large increases in development in the lower coastal plain (LCP) regions of the state over the next 30 years (Allen and Lu, 2003). While development is good for the economic interests of an area, it is often accompanied by land cover changes that have adverse effects on local watersheds. Impervious cover typically increases when land development takes place, and the percentage of impervious cover in a watershed is now used as an environmental indicator to assist local authorities with resource protection planning (Arnold and Gibbons, 1996). Land cover change associated with land development changes runoff patterns and alters local streamflows which can result in declining biological health in developing watersheds (Booth et al., 2002). Water quality often declines in developing watersheds as well due to increasing non-point source runoff which has been specifically demonstrated in the sensitive tidal marsh ecosystems in the LCP (Holland et al., 2004). With typical development and subsequent impervious cover increase, storm runoff will increase in volume and peak discharge, altering pre-development site hydrology and creating issues downstream (Blair et al., 2011). In order to accommodate increasing development in the LCP, adequate stormwater management measures must accompany land cover changes to ensure the continued health of area watersheds. The performance of such stormwater handling measures depends on data-driven design that accounts for the expected rainfall response for area watersheds based on pre-development runoff generation. This situation is complicated in the LCP because of unique hydrologic conditions that differ from the better studied hydrology of upland watersheds.

Lower Coastal Plain Hydrology

Coastal headwater streams in undeveloped forested landscapes function as a natural storage and conveyance mechanism for groundwater discharges and streamflow (Amatya et al., 2006). The LCP of South Carolina is defined by low gradient topography and low elevations typical of southeastern United States coastal landscapes. Shallow groundwater elevations influence soil moisture levels and couple with surface water generation during rainfall events to determine stream outflows that include significant baseflow (BF) (Eshleman et al., 1994; Williams, 2007). The magnitude of watershed outflows is often driven by a fluctuating water table position that is regulated by the balance between evapotranspirative demand and infiltrative replenishment by rainfall (Miwa et al., 2003; Amatya et al., 2006; Slattery et al., 2006; Harder et al., 2007). High water table elevation and high soil moisture conditions lead to higher outflow production during winter months when forest vegetation is largely dormant and evapotranspiration rates are lower (Harder et al., 2007; Williams, 2007; Amatya and Skaggs, 2011; La Torre Torres et al., 2011). During summer months, streamflows are intermittent in response to direct rainfall. High summer evapotranspiration rates tend to rapidly lower the water table elevation resulting in increased soil storage and decreased storm runoff (Slattery et al., 2006; Harder et al., 2007; Williams, 2007; Amatya and Skaggs, 2011). Flow cessation occurs when the water table elevation is sufficiently low that groundwater flows are disconnected from the stream channel. Between these seasonal extremes, the rainfall response is dependent upon antecedent moisture conditions (AMC) that vary with microclimate variability and seasonal evapotranspiration shifts (Sun et al., 2002; Harder et al., 2007). Due to these highly variable conditions, the derivation of water budgets for coastal forested watersheds with low-gradient topographic relief can be complex.

Though much work has been done to characterize upland watersheds, more information is needed specific to the LCP with respect to regional hydrologic processes (Amatya et al., 2006). Previous studies of headwater catchments in the LCP have reported variable annual outflows as a percentage of rainfall. Amatya et al. (2006) measured total annual outflow depth as a percentage of rainfall over a long-term dataset covering 23 years on two first-order forested watersheds in the Francis Marion National Forest located in the LCP. Results ranged from 5% to 59% on a control watershed (WS80) and from 9% to 44% on a treatment watershed (WS77). Differences in total outflow between years are due to variations both in the temporal distribution of annual rainfall and the AMC at the time of rain. In coastal forest water budgets, the relationship between rainfall and outflow production is affected by soil moisture levels that are influenced by the shallow water table (Amatya and Skaggs, 2011). In a study of a drained pine plantation in eastern North Carolina, Amatya and Skaggs (2011) highlighted seasonal differences in moisture levels that resulted in wet conditions and higher outflows during winter months and dry conditions and intermittent outflows during summer months. This seasonal trend was related to variable water table elevations that produced outflows by subsurface flow when within 1.1 m of the ground surface. Harder et al. (2007) computed water budgets for WS80 over two consecutive years and measured a total outflow depth as a percentage of rainfall of 0.47 for 2003 and 0.08 for 2004. This range in outflows between years was partially due to differences in annual rainfall (1670 and 960 mm, respectively) and partially due to differences in AMC. Several large storms during 2004 resulted in only moderate to low outflows, and this was linked to lower water table elevations at the time of rainfall that characterize dry AMC. This was also shown by Dai et al. (2011) in their modeling study. Using bi-criteria of streamflow and water table depth in the distributed hydrological model MIKE SHE, better results in calibration and validation of model parameters were obtained for a LCP headwater stream than were found using the single criterion model with just streamflow. This demonstrates the role that water table elevation has in determining soil moisture conditions and streamflow on these watersheds. In another study, Sun et al. (2002) compared the hydrologic response of two flat LCP watersheds in North Carolina (NC) and Florida (FL) to a high gradient watershed with considerable topographic relief in the Appalachian region of North Carolina (UP) using long-term precipitation and flow data. Using the ratio of annual outflow depth to annual rainfall, the UP watershed demonstrated higher results (0.53) compared to the two lowland watersheds (0.30 for NC, 0.13 for FL). Climate variability was one factor for the difference, with average annual precipitation of 1730 mm for the UP watershed and 1520 mm and 1260 mm for the LCP watersheds in NC and FL, respectively. Outflow in the high-gradient watershed had consistent BF contributions and flowed constantly during the study period. Intermittent flow was observed on the LCP streams reflecting variable water table elevation and intermittent groundwater discharges. Variable AMC affects BF levels as these conditions change and stream outflows behave accordingly. A study by Eshleman et al. (1994) observed that sustained BF contributions on low gradient coastal plain watersheds are almost entirely subsurface groundwater discharges. A chemical tracer study revealed that storm flow was composed mostly of "old" water contributions from groundwater discharge. "New" water contributions were the result of direct channel interception of rainfall and saturated overland flow in riparian areas of the watershed with high soil moisture that varied in size according to moisture conditions. These observations are inconsistent with Todd et al. (2006) who reported that for wetland-dominated watersheds, hydrologic processes are inconsistent with some aspects of the variable source area concept of streamflow generation. The authors found that some parts of the basin may become decoupled from the basin outlet as summer progresses. Runoff from these portions of the watershed may be lost to evaporation and infiltration or held in surface storage even before reaching the outlet.

Outflows on these LCP headwater streams are not determined by rainfall alone because of variable AMC consistent with Todd et al. (2006) who noted that the vertical water movement due to rainfall, ET, and deep seepage was more important than the lateral groundwater flux in explaining the wetland's hydrologic behavior. Harder et al. (2007) showed that the temporal distribution of rainfall as it relates to AMC has more influence on outflow production than just rainfall totals alone. Inspection of storm events during the summer of an overall dry year revealed substantial storms that resulted in moderate to no outflows. Sun et al. (2002) analyzed total outflows related to isolated storm events on the LCP FL watershed. Storm events were selected to assess the difference in outflows generated by both small and large storms that fell on the watersheds for both dry and wet AMC. The magnitude of streamflow prior to rainfall, relative to typical outflows on the watershed, was used to differentiate between dry and wet AMC. Small storms in the range of 30 – 59 mm demonstrated lower storm outflow depths on the FL watershed that were consistent with lower annual outflows. Large storms in the range of 102 – 160 mm demonstrated lower storm event outflows at the FL watershed for dry AMC (0.08 as a ratio of event rainfall) and much higher storm outflow depth for wet AMC (0.58 as a ratio of event rainfall). The large increase in outflow production in the rainfall response from dry AMC to wet AMC at FL demonstrates the role of soil storage on outflow production on these LCP headwater streams. The dry AMC is associated with lower water table elevations and higher soil storage that is filled before runoff is generated. The wet AMC is characterized by high water table elevations with low soil storage, and these conditions generate runoff rapidly under saturated conditions (Amatya and Skaggs, 2011).

Storm event outflow production has been studied in order to determine the role that seasonal trends in evapotranspiration, water table elevation, and AMC have on LCP headwater streams in the rainfall response. A study by La Torre Torres et al. (2011) demonstrated that storm flows change seasonally in step with trends in AMC. Storms selected from a long-term dataset at Watershed 78 in the Francis Marion National Forest in the LCP of South Carolina were separated according to the wet (Dec. – May) and dry seasons (June – Nov.). Results demonstrated a significant relationship between rainfall and storm flows (total event outflow as a percentage of rainfall) during the wet season ($r^2 = 0.68$, $p < 0.01$) and less so during the dry season ($r^2 = 0.19$, $p = 0.02$). Rainfall accounts for a lower amount of variability in storm flows during the dry months due to lower water table elevation and higher soil storage caused by increased ET demands. The authors also suggested that the storm flows were controlled mainly by rainfall amount and the AMC represented by the initial flow rate. In a coastal North Carolina study, Amatya et al. (2000) found the drainage outflow as a percentage of rainfall varying between 2% for a dry summer period with water table as deep as 2.3 m to as much as 89% for a wet winter event with near surface water table for a drained pine forested watershed. These authors also reported that the amount of rainfall needed to generate a drainage event depends upon the AMC which was represented by initial flow at the watershed outlet. Williams (2007) observed a non-linear relationship between rainfall and direct runoff estimates in an LCP headwater stream. Runoff generation mechanisms unique to LCP hydrology were related to water table elevation, and runoff production was observed to increase sharply when the water table elevation was within 1.0 m of the ground surface at the watershed divide. BF levels were modeled by a graphical hydrograph separation method that accounts for accelerated groundwater discharges in the rainfall response. These flows were found to track groundwater elevations closely due to the relationship between water table elevation and watershed outflows for these streams. An investigation by Rogers et al. (2009) of the rainfall response on a LCP headwater stream at Upper Debidue Creek (UDC) near Georgetown, SC showed that groundwater elevations tracked the stream outflow hydrograph closely. Direct runoff was estimated by hydrograph separation for storm events and increased with consecutive closely spaced storms that contributed to increasingly wet watershed conditions and higher runoff generation (Rogers et al., 2009). The assessment of the rainfall response is complicated by the interaction between groundwater discharges and surface water generation on LCP watersheds. Separation of storm flow into BF and surface generated runoff components is difficult due to this interaction. Water table dynamics that are influenced by climate and evapotranspiration contribute to variable AMC on LCP headwater streams. This results in a range of outflows as a percentage of rainfall on an annual basis as well as between storm events. Most recently, Callahan et al. (2011) estimated total recharge to groundwater for the same LCP watershed studied by La Torre Torres et al. (2011) by analyzing water table response to storm events and the rate at which water was transferred into the shallow aquifer. The authors attributed the difference found in two methods of estimating the recharge (water table fluctuation method and Darcy equation) to the ET that took place and AMC prior to the rain event that were not accounted for in the Darcy method. Variable moisture conditions linked to seasonal trends in ET and water table elevation create a variable response to rainfall on LCP headwater streams, and increased knowledge of these baseline hydrologic conditions will help guide better watershed protection measures as development increases in the area.

Canopy Interception and Throughfall

Tree canopy and vegetation surfaces in a forested watershed act as an initial barrier to rainfall. Only part of the rain that falls in a forested watershed passes through the canopy as throughfall. The portion that is intercepted by vegetation is either funneled toward the ground as stemflow or it evaporates back into the atmosphere as canopy interception, never entering the watershed. This canopy interception portion of rainfall has been studied in length because of its ability to alter the quantity, timing, and areal distribution of incoming precipitation to a catchment (Swank, 1968). It is an important component of the forest water budget that has shown to account for up to 35% of annual rainfall in forested watersheds of the southeastern United States (Swank, 1968; McCarthy et al., 1991). The amount of intercepted rainfall is influenced by physical canopy characteristics as well as climatological conditions. Swank (1968) showed that the manipulation of forest cover from hardwood species to conifers substantially lessened the amount of rainfall entering the watershed and subsequent outflows were lower under cover of the pines. This change in canopy cover resulted in an altered water budget that is attributed to differences interception due to the physical structure of the two types of trees. Comparison of similar forest stands in the piedmont and coastal regions of South Carolina in the study highlighted climatological differences that impacted canopy interception as well. Rainfall patterns and storm size played a role in the differing levels of interception between similar stands in the two geographical areas. Canopy interception differs in amount based upon rainfall characteristics and the physical structure of watershed vegetation. The spatial variability of these factors (rainfall and canopy cover) makes it a difficult component of the water budget to measure.

In a summary of previous studies on canopy interception, Crockford and Richardson (2000) identified the main sources of variability in canopy interception as factors of forest type or climate. When rain falls on the forest canopy, it is partitioned into one of three fractions. Throughfall is the portion of the rain that falls to the ground, stemflow is that which is collected by canopy structures and funneled to the ground separately, and interception is the portion stored on vegetation surfaces and evaporated back into the atmosphere. This partitioning is a direct result of the physical structure of the canopy itself. McCarthy et al. (1991) identified canopy closure, leaf area index (LAI), and canopy storage capacity as the most important physical factors of canopy structure that determine interception. These are measurements of canopy density and vegetation surface area and they provide a good measure of how likely the vegetation is to intercept and store incoming rainfall. Climatological factors identified by Crockford and Richardson (2000) are characteristics of the rainfall itself (amount, intensity, and duration) and other factors that influence evaporation rates of intercepted rainfall (wind speed, temperature, and humidity). Rainfall that is intercepted is eventually evaporated back to the atmosphere. The process of rainfall deposition on vegetation surfaces and evaporation may take place more than once during any given rain event depending on climate factors that influence evaporation. The canopy interception process can thus be modeled based on the ability of the forest canopy to collect and store rainfall and the climatological factors present that inhibit or accelerate the evaporation process. Canopy interception is difficult to measure on the landscape level because of its large spatial variability (Crockford and Richardson, 2000). Tree structure at the individual level affects this process, and structure can vary greatly depending on species and age (Swank et al., 1972). Structural differences between species affect how much rain is intercepted, stored, evaporated, or funneled to the ground as stemflow. Interception amounts are dependent on the organization of biomass of the species in both lateral spread and vertical composition (Lefsky et al., 1999). This is most noticeable between conifers and deciduous trees, which have very different shape, branch structure, and leafing patterns. Interception differences between species vary on a seasonal level due to leaf loss in deciduous species as well (Helvey and Patric, 1965b). Common interception levels can be found within certain forest types, but even this varies geographically due to differences in climate (Crockford and Richardson, 2000). Spatial differences in interception create difficulty in reliable direct measurement.

Canopy interception (I, mm) is typically measured as the difference between gross precipitation (P, mm) and the portion of rainfall that enters the watershed, throughfall (T, mm) and stemflow (S, mm), and is represented by the following equation:

$$I = P - (T + S) + L \tag{2-1}$$

Interception losses from leaf litter (L, mm) are included here but most studies omit them due to difficulty in measurement, high variability, and small contribution (Helvey and Patric, 1965a). Interception is rarely measured directly and it is appropriate to consider the process in terms of the effective rainfall (R, mm), defined as the net precipitation entering the soil and contributing to the water budget (Swank, 1968).

$$R = P - I = T + S - L \tag{2-2}$$

Some studies omit stemflow measurement as well as litter interception due to insignificant levels and difficulty in measurement. Variability in the measurement of leaf litter interception and stemflow and the low contributions of both (on the order of 0 – 9% of gross rainfall for stemflow and 2-5% for leaf litter interception (Helvey and Patric, 1965b; Helvey, 1971; Swank et al., 1972)) can have a cancelling effect. The majority of canopy interception is largely driven by canopy storage and evaporation, and this can be approximated passively by the measurement of throughfall under the canopy.

The throughfall and stemflow portions of incoming rainfall are typically measured by random placement of multiple rain gages under the canopy for throughfall and tree collars for stemflow to account for and minimize the effects of spatial variability in these processes (Calder and Rosier, 1976; Crockford and Richardson, 2000). The lack of a standard practice for gage type, number of gages, or placement within the watershed in interception studies has led to results that are difficult to extrapolate outside of the specific conditions studied due to the sources for error and variability between canopy and climate at different locations (Helvey and Patric, 1965a). Helvey and Patric (1965a) attempted to measure the variability in measurements of different portions of canopy interception through analysis of the covariance over a number of studies. Levels of variability in measurement were applied to develop recommendations for adequate sampling for studies to use to aid in the comparison of results. Rigorous statistical assessment of a study is necessary to determine adequate sampling procedures to ensure confidence and precision in water budget calculations that can be compared to results from different sites (Zarnoch et al., 2002). Adequate sampling may still produce inaccurate measures of throughfall though as values greater than gross precipitation have been reported (Crockford and Richardson, 2000). Many research efforts have focused on the development of a model that is able to accurately account for these physical and spatial variables and produce reliable estimates for canopy interception across all conditions due to the rigorous measurements needed to obtain acceptable accuracy. These models focus on the physical processes of canopy interception and vary greatly in simplicity, accuracy, and applicability.

The most widely used models of interception have been empirical regressions between interception (I) and gross precipitation (P) of the form:

$$I = aP + b \tag{2-3}$$

In this model, *a* and *b* are regression coefficients obtained through linear regression (Gash, 1979). Swank (1968) traces the development of these linear regression equations back to the work of Horton (1919), who quantified the physical process of interception loss (I, mm) as

$$I = S_j + K_1 E_r T_s \tag{2-4}$$

where S_j is interception storage capacity (mm), K_1 is the ratio of evaporative surface over the projectional area, E_r is the evaporation rate (mm/hr) during the storm, and T_s is the duration of the storm (hr). Swank (1968) goes on to describe the work of Kittredge (1948), who rewrote the equation to include the influence of gross precipitation (P, mm) as

$$I = S_j + (K_1 E_r T_s / P)P \tag{2-5}$$

This interpretation apportions interception into a canopy storage component and an evaporation component. If the evaporation component during any given storm is assumed to be a constant proportion of rainfall, interception becomes a linear function of gross precipitation represented by Equation 2-3. Substitution of Equation 2-3 for interception into Equation 2-2 for R (and assuming that S and L are negligible) produces a linear expression for throughfall (T) as a function of P.

$$R = T = P - I = P - (aP + b) = (1-a)P - b \tag{2-6}$$

This linear expression for T approximates the portion of rainfall not lost to evaporation (1-a) that is a function of P, decremented by a measure of canopy storage (b). This simplistic approach for modeling interception by linear regression of measured P and T has been applied to many forest stands in different areas in the eastern and southeastern United States alone (summarized in Table 2.1).

Table 2.1. Summary of throughfall regression equations for studies on tree species/ communities common to the lower coastal plain of South Carolina.

Study	Location	Vegetation Type	Throughfall Regression
Blood et al., 1986	Coastal South Carolina	Loblolly pine (*Pinus taeda*)	0.76P-0.06
Helvey, 1971	Multiple	Loblolly pine (*Pinus taeda*)	0.80P-0.01
	Mutiple	Average Pines	0.86P-0.04
Helvey and Patric, 1965b	Eastern United States	Hardwoods (Growing Season)	0.90P-0.03
		Hardwoods (Dormant Season)	0.91P-0.02
Roth and Chang, 1981	East Texas	Longleaf pine (*Pinus palustris*)	0.90P+0.10
Swank et al., 1972	Piedmont South Carolina	Loblolly pine (*Pinus taeda*), 5 yr.	0.83P-0.03
		Loblolly pine (*Pinus taeda*), 10 yr.	0.73P+0.00
		Loblolly pine (*Pinus taeda*), 20 yr.	0.76P+0.01
		Loblolly pine (*Pinus taeda*), 30 yr.	0.85P+0.00
		Hardwood/pine mix, mature	0.87P-0.02

These studies characterize the difference in throughfall for stands of different forest communities, species type, age within species type, silvicultural treatment, and land cover change. Collectively they display the variability in regression equations that is attributed to differences in stand properties and the influence of physical structure on canopy interception.

Greater accuracy in modeling canopy interception has been sought to refine water budget calculations, and much of this work has focused on the inclusion of meteorological parameters. These parameters are known to affect evaporation levels during a storm and impact rainfall contributions to filling the canopy storage capacity.

Rutters et al. (1975) developed a model that conducts a water budget on an hourly basis utilizing meteorological inputs and it has performed well against measured data. This model requires hourly meteorological measurements, which are often not available, and a complex program to operate. Because of this, Gash (1979) developed a model that incorporated more physical parameters into the linear model to create a more accurate method for interception estimation that honors the physical drivers of the process and maintains simplicity. In the model, interception is separated into time components representing the process up to and after the canopy storage is filled. Before the storage capacity is filled, interception is a function of the time it takes the canopy to reach this capacity subject to concurrent evaporation and the density of the canopy itself. Once storage capacity is reached, remaining interception is a function of evaporation rates. This is modeled as

$$I = (\bar{E}/\bar{R})P_G + \left(S + \int_0^{t'} E\,dt\right)\{1 - (\bar{E}/\bar{R})(1 - p - p_t)^{-1}\} \qquad (2\text{-}7)$$

where $\overline{E}$ is the mean evaporation rate (mm/hr), $\overline{R}$ is the mean rainfall rate (mm/hr), S is the canopy storage capacity (mm), $\int_0^{t'} E\,dt$ is the evaporation up to the time canopy storage is filled (t'), p is the free throughfall coefficient, and p_t is the proportion of rainfall diverted to streamflow. Examination of this equation reveals that it approximates the linear regression equation and provides a physical basis for the regression coefficients of *a* and *b* in Equation 2-3. Even so, this model is data intensive and subject to the same variability of canopy and climatological factors that make extrapolation of results between sites unreliable. Development of rapid spatial canopy assessment technology will strengthen the modeling of accurate canopy interception at a much more broad geographic level.

Determining the Rainfall Response

Towards the goal of determining the response to rainfall for a watershed of study, total storm flows do not always characterize the rainfall response well. Base flows (BF) and subsurface contributions of groundwater to streamflows during rain events can influence storm flows and do not relate to surface-driven runoff generation. Storm event hydrographs are typically apportioned between components of flow that link a certain percentage of the total storm flow to a specific source in the watershed. This is of increasing concern as land cover changes have been linked to changes in surface generated stormwater runoff, and estimating the direct runoff (DR) portion of streamflow has become important. In a study of smaller forest watersheds, Hewlett and Hibbert (1967) used runoff generation mechanisms on forested watersheds to better advise the apportionment of streamflow between the components of channel interception, subsurface interflow, DR, and BF. The difficulty in reliable apportionment between these components of streamflow is linked to varying site conditions for any given storm event that contributed to a variable response to rainfall. The authors proposed the concept of variable source areas defined by changing moisture conditions on a watershed that contribute to variation in runoff generation for storm events. These variable source areas account for differences in runoff generation between similar storms and represent the physical processes of runoff generation on forested watersheds better than previous models. This concept is valuable in explaining the range of runoff generation for any given watershed and it highlights the difficulty in modeling the rainfall response. A range of hydrograph separation methods have been developed to separate storm event hydrographs into BF and DR components to better measure the rainfall response.

In a review of techniques for separating BF components from the streamflow hydrograph, Brodie and Hostetler (2005) identified techniques that could be classified into the three categories of BF separation, frequency analysis, and recession analysis. BF was identified as the more long term delayed flow from watershed storage and the remaining portion of streamflow was termed “quick flow,” representing the immediate storm response of the watershed and equivalent to DR. BF separation methods are most often graphical methods of hydrograph separation that use analysis of time-series data for streamflow to distinguish between BF and DR components. These methods employ different algorithms or decision equations to define the start and end of DR contributions to streamflow (Chapman, 1999). BF is then defined between these points according to known hydrologic processes for a given site that relate groundwater discharges to streamflow production. This can be as simple as a straight line connecting the start and end of DR to more complex models that relate BF to storage-discharge relationships of the groundwater aquifer that are linear or even quadratic (Chapman, 1999). The second category of hydrograph separation methods is frequency analysis. This is the least studied and involves flow duration curves that describe how frequently a flow rate is equaled or exceeded. Analysis of these curves has yielded conclusions about BF levels, but work on this is ongoing. The third category of hydrograph separation methods is recession analysis which involves only the receding limb of the hydrograph after peak flow has passed. Different segments of the recession limb are analyzed, often graphically, to characterize outflow decay in order to discern the storage-discharge relationship of BF contributions. This relationship is then used to separate the recession limb between BF and DR. In a study of recession curve analysis techniques, Tallaksen (1995) discussed the complexity of recession modeling due to the variability of controlling factors on different watersheds. A seasonal difference was identified that was related to seasonal changes in evapotranspiration and linked to faster recession rates for hydrographs of storm events during summer months. The wide variety of hydrograph separation techniques yield different results for BF and DR estimates and it is necessary to choose a method that accounts for site specific characteristics of BF discharges and storm response runoff generation.

Distinction between BF and DR in the LCP is complicated by interactions between surface water and groundwater during the rainfall response. A study of groundwater discharges and runoff generation was conducted by Eshleman et al. (1994) on a coastal plain watershed in Virginia that had similar conditions to the LCP of South Carolina. Tests of sustained BF and the saturated hydraulic conductivity of the groundwater aquifer revealed that streamflows outside of the rainfall response consisted solely of shallow groundwater discharges. Chemical tracers were used to determine the source of streamflows during storm events, and it was determined that “old” water dominated storm flows. The highest contributions of “new” water (rainfall) to storm flows occurred at the peak of the hydrograph and measured up to 40% of streamflows. The total volume of “new” water for a given storm event was divided by the rainfall depth to approximate the area of the watershed contributing this rainfall depth directly to storm flows by channel interception or saturated overland flow. This area was comparable to estimates of saturated areas for the storm events measured. This supports variable source runoff generation by saturation excess overland flow in saturated riparian areas as the dominant mechanism in low gradient watersheds found in the coastal plain. In a study to model conditions contributing to the formation of variable source areas in humid watersheds with shallow water table, Hernandez et al. (2003) determined that mild slope, shallow water table, and low hydraulic conductivity were the most amenable conditions to these variable source areas of runoff. Models used digital elevation models and a variable saturation model combined with rainfall characteristics and soil moisture conditions to assess the formation and morphology of the variable source areas near wetlands and streams. Low gradient and shallow water table in LCP watersheds likely contribute to these conditions and may explain the variable rainfall response. A measure of the rainfall response often used for storm event analysis is the runoff coefficient (ROC), defined as DR contributions as a depth divided by storm event rainfall. This represents the portion of rainfall that exits as quick response storm flow. A study of ROCs measured from long-term datasets for a large number of sites revealed that the ROC ranged according to moisture levels on the watershed prior to the storm event as measured by soil moisture or base flow before the start of rainfall (Longobardi et al., 2003). The rainfall response for these watersheds was described and modeled as the summation of two linear reservoirs discharging at different rates. The first represents BF from groundwater discharges and the outflows are longer and more delayed. The second discharges quicker in response to rainfall and represents surface flows and shallow subsurface contributions to streamflow and are assigned as DR. Williams (2007) also modeled the storm response according to the parallel linear reservoirs discharging at different rates. Williams uses the linear model of Maillet (1905) to model BF recession. This model describes outflow rate at a particular time as a function of initial flow rate, time and a rate constant. The model is expressed as:

$$Q_t = Q_0 * \exp(-t/k) \qquad (2\text{-}8)$$

where Q_t is the outflow rate at time t, Q_0 is the outflow rate at the start of the recession, and k is a rate constant. This model was used to model BF from the hydrograph peak until the end of the hydrograph. A straight line from the initial hydrograph rise to the peak BF value was used to model increasing BF contributions to streamflows during the rising limb of the hydrograph. Rapid water table response to rainfall at the watershed divide in LCP forested watersheds was observed to increase the hydraulic gradient toward the stream channel and increase groundwater discharges soon after rainfall began. Streamflows above this separation line were assumed to be DR. The interaction of groundwater and surface water generation on LCP watersheds must be recognized and accounted for in order to determine the rainfall response in these watersheds. BF contributions in the rainfall response may be more of a function of groundwater elevation, and groundwater elevation has been shown to contribute to surface runoff generation on these low gradient watersheds.

Curve Number Method

The CN method (as described in USDA Technical Release 55 (TR-55), 1986) assigns CNs to land surfaces based on hydrologic soil group (HSG), cover type, treatment, hydrologic condition, and antecedent runoff condition (ARC). The CN is a function of maximum potential retention, and it can be interpreted as the degree of permeability for the corresponding land cover conditions. Curve numbers have a range from 0 – 100 representing conditions from infinite infiltration to fully impermeable, respectively. Typical observed values range from 40-98, however, though they may be lower for forested conditions (Van Mullem et al., 2002). Soil types are assigned to one of four hydrologic soil groups (A, B, C, or D) based on infiltration and hydraulic conductance properties, with A being the most permeable and D the most impervious. Wet soils are assigned dual HSGs (A/D, B/D, C/D). These soils are assigned as Group D in the undrained condition and are better modeled as the alternate HSG if adequately drained (USDA, 2007). Typical land cover types have been classified, and a range of CNs for each has been calculated across the spectrum of HSGs. The cover types are further broken down by treatment (method of land management) and hydrologic condition (Good, Fair, or Poor; based on runoff potential and typically measured by density of plant cover) where applicable. ARC is a measure of antecedent moisture and it accounts for the range in runoff response that can be expected from dry (CN-I) to wet (CN-III) conditions. Most CN applications use the average ARC (CN-II) for runoff estimates.

The CN is a transformation of the variable S (mm), which represents the potential maximum retention of rainfall by the land.

$$CN = 25{,}400/(S - 254) \qquad (2\text{-}9)$$

S can also be thought of as the greatest possible difference between rainfall (P, mm) and DR (Q, mm) for any given storm event. Representative CNs for the combination of land cover conditions and soil composition for a site are weighted by respective area percentages to produce a composite CN. This CN is applied to the CN equation to predict the direct runoff to be expected from any given storm.

$$Q = \frac{(P-I_a)^2}{(P-I_a)+S} \text{ where } P > I_a\text{, otherwise } Q = 0 \qquad (2\text{-}10)$$

The remaining variable in the CN method is I_a (mm), or the initial abstraction. This variable represents the portion of rainfall that does not produce DR. Initial abstraction is a composite of canopy interception, infiltration, surface storage, and other losses deducted from rainfall before DR is produced (USDA, 1986). The quantity ($P - I_a$) is equivalent to the effective precipitation producing runoff for a storm event. The initial abstraction was originally set at 20% of S based on calibrations performed in the development of the model. This simplified the CN method to one independent parameter, P, once the site CN was defined.

$$Q = \frac{(P-0.2S)^2}{(P+0.8S)} \text{ where P > 0.2S, otherwise Q = 0} \qquad (2\text{-}11)$$

Once the CN for a site has been measured by land cover and soils analysis, Equation 2-11 can be used to predict the runoff depth for any given rainfall.

Use of the CN method for runoff prediction applications has been questioned in some areas due to its wide use for a variety of hydrologic conditions that were not considered in its development (Ponce and Hawkins, 1996). Proper model parameter selection is crucial for reliable estimates of DR, but even this may not produce realistic results for some hydrologic conditions due to the assumptions that the model makes in regards to runoff generation. Though the model is simple and generally reliable for watershed outflow prediction in many cases, it often lacks true representation of the physical processes involved in runoff generation (Boughton, 1989). Representation of different hydrologic conditions can be accomplished by varying the parameters of the model to more accurately reflect watershed characteristics and conditions. This is rarely conducted for the sake of simplicity despite notice in TR-55 of site-specific deviation from typical CN method applications (Ponce and Hawkins, 1996).

Improper selection of site CN can be an initial source of error in using the CN method. DR estimates are more sensitive to the CN than other parameters. For examples, Boughton (1989) has shown that a 15-20% change in CN almost doubles or halves DR predictions. Curve number selection involves the classification of site conditions by discrete categories as defined by TR-55. Natural deviation among these conditions and the potential for misclassification may produce unrealistic runoff estimates due to incorrect CN selection, especially for forested watersheds (Hawkins, 1993). Dual HSG soils further complicate CN selection, as the additional site classification between drained and undrained produces a large difference in runoff estimates. Discrete measurement of site CN is difficult because runoff production is variable between storm events. The CN has been interpreted as a random variable that ranges for any given storm based on the ARC (Hjelmfelt, 1991; Van Mullem et al., 2002). The CN method offers very little guidance on accounting for differences in runoff production between dry and wet conditions. ARC was initially based on 5-day antecedent precipitation index (API), but this was later revised due to differences in regional definitions for site moisture (Ponce and Hawkins, 1996). No specific CN method guidelines for ARC determination are currently offered in TR-55. CN tables for site determination are listed for the average ARC (CN-II), interpreted as the median CN measured by analysis of rainfall and runoff data. A correction must be applied to the CN-II for the dry ARC (CN-I) and wet ARC (CN-III) (Ponce and Hawkins, 1996). These values are considered probabilistic upper and lower limits for runoff production for a given site based on the range of soil moisture conditions (Hjelmfelt, 1991; Ponce and Hawkins, 1996). Accounting for differences in runoff production according to ARC is based upon user discretion and subsequent parameter adjustments. Selection of the CN that is best suited to a given watershed can be difficult and prone to error.

The initial abstraction was originally set at 20% of S to simplify the CN method to one parameter. This does not account for differences in site conditions, hydrology, or runoff generation mechanisms. This parameter determines the effective precipitation that contributes to direct runoff production. Woodward et al. (2003) showed that a level of 5% is more realistic by comparing data from over 300 watersheds over the eastern two-thirds of the United States. Lim et al. (2006) supported this lower measure with GIS modeling in a watershed in the Midwestern United States. Despite this, adoption of a median value for initial abstraction for all CN applications does not reflect regional watershed differences in runoff production. Ponce and Hawkins (1996) suggested from the work of Bosznay (1989) and Ramasastri and Seth (1985) that initial abstraction might be better interpreted as a regional parameter that reflects differences in geography and climate for better model results. TR-55 notes that care should be taken to ensure that the assumptions made in using this initial abstraction term reflect field measurements. The use of a different value for initial abstraction requires changes in CN tables due to associated changes in CN equations. Widespread use of the CN method is a function of the relative simplicity and reliability of the model, but sources for error in parameter estimation should be considered for model refinements.

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CHAPTER THREE

SPATIAL VARIABILITY IN RAINFALL AND THROUGHFALL MEASUREMENT

"Now the storm is over, the sky is clear, the last rolling thunder-wave is spent on the peaks, and where are the raindrops now – what has become of all the shining throng? In winged vapors rising some are already hastening back to the sky, some have gone into the plants, creeping through invisible doors into the round rooms of cells, some are locked in crystals of ice, some in rock crystals, some in porous moraines to keep their small springs flowing, some have gone journeying on in the rivers to join the larger raindrop of the ocean." – John Muir, *My First Summer in the Sierra*

ABSTRACT

Rainfall and subcanopy throughfall were measured over a year at Upper Debidue Creek (UDC) to assess spatial variability and throughfall levels. Hourly rainfall measurements were analyzed at five open field rain gages to determine if there was spatial variation that requires a network of rainfall measurement by multiple gages. It was determined that variation in hourly measurements at the additional rain gages were within manufacturer's accuracy limits for the rain gages. The open field rain gage that has been used at UDC is sufficient to capture hourly rainfall totals for the watershed. Throughfall measurements at ten static subcanopy gages do not measure spatial variations in canopy throughfall well. Seasonal variation in throughfall measurement was observed that reflects the role of additional canopy density and higher evapotranspiration

during the growing season. A linear regression for expected throughfall for the watershed was calculated from previous literature studies of throughfall and the vegetation composition of the UDC watershed that can be used to approximate throughfall from open field rainfall measurements.

BACKGROUND

Tree canopy and vegetation surfaces in a forested watershed act as an initial barrier to rainfall. Only part of the rain that falls in a forested watershed passes through the canopy as throughfall. The portion that is intercepted by vegetation is either funneled toward the ground as stemflow, or else it evaporates back into the atmosphere as canopy interception and thus never enters the watershed. The canopy interception portion of rainfall has been studied in length because of its ability to alter the quantity, timing, and areal distribution of incoming precipitation to a catchment (Swank, 1968). Canopy interception is an important component of the forest water budget that accounts for up to 35% of annual rainfall in forested watersheds of the southeastern United States (Swank, 1968; McCarthy et al., 1991). The amount of intercepted rainfall is influenced by physical canopy characteristics as well as climatological conditions. Swank (1968) showed that the manipulation of forest cover from hardwood species to conifers substantially lessened the amount of rainfall entering the watershed and subsequent outflows were lower under cover of the pines. This change in canopy cover resulted in an altered water budget that is attributed to differences in the physical structure of the two types of trees. Comparison of similar forest stands in the piedmont and coastal regions of South Carolina in the study highlighted climatological differences. Rainfall patterns played a role in the differing levels of interception between the two geographical areas. Canopy interception differs in amount based upon rainfall characteristics and the physical structure of watershed vegetation. The spatial variability of these factors (rainfall and canopy cover) makes it a difficult component of the water budget to measure.

Water budgets in the Lower Coastal Plain (LCP) of South Carolina have been studied in detail in order to establish a baseline for watershed dynamics to inform better decision making as land cover change increases in the area. Low gradient and a seasonally shallow water table create variable outflows over the course of the year and in response to rainfall. The differences in these watersheds from those more typical in higher gradient landscapes have been characterized by several studies in the LCP to highlight these differences (Sun et al., 2002; Amatya et al., 2006; Harder et al., 2007; Williams, 2007; La Torre Torres et al., 2011). In order to refine water budgets and quantify the rainfall response, it is necessary to assess canopy interception levels on undeveloped watersheds in the LCP. Streamflows in headwater catchments in the LCP are typically intermittent during the summer months due to seasonally low groundwater elevations. Sustained streamflows are typical during the winter months when groundwater elevations are higher. This trend in groundwater elevation is linked to seasonal shifts in evapotranspiration (ET). Canopy interception depends on vegetation density and evaporation and a relationship between interception levels and seasonal shifts in ET is expected. Measurements of interception will provide insight into rainfall response dynamics and the potential impacts on water budgets as land cover changes reduce canopy coverage in LCP watersheds.

In a summary of previous studies on canopy interception, Crockford and Richardson (2000) identified the main sources of variability in canopy interception as factors of forest type or climate. When rain falls on the forest canopy, it is partitioned into one of three fractions. Throughfall is the portion of the rain that falls to the ground, stemflow is that which is collected by canopy structures and funneled to the ground separately, and interception is the portion stored on vegetation surfaces and evaporated back into the atmosphere. This partitioning is a direct result of the physical structure of the canopy itself. McCarthy et al. (1991) identified canopy closure, leaf area index (LAI), and canopy storage capacity as the most important physical factors of canopy structure that determine interception. These are measurements of canopy density and vegetation surface area and they provide a good measure of how likely the vegetation is to intercept and store incoming rainfall. Climatological factors identified by Crockford and Richardson (2000) are characteristics of the rainfall itself (amount, intensity, and duration) and other factors that influence evaporation rates of intercepted rainfall (wind speed, temperature, and humidity). Rainfall that is intercepted is eventually evaporated back to the atmosphere. The process of rainfall deposition on vegetation surfaces and evaporation may take place more than once during any given rain event depending on climate factors that influence evaporation. The canopy interception process can thus be modeled based on the ability of the forest canopy to collect and store rainfall and the climatological factors present that inhibit or accelerate the evaporation process. Canopy interception is difficult to measure on the landscape level because of its large spatial variability (Crockford and Richardson, 2000). Tree structure at the individual level affects this process, and structure can vary greatly depending on species and age (Swank et al., 1972). Structural differences between species affect how much rain is intercepted, stored, evaporated, or funneled to the ground as stemflow. Interception amounts are dependent on the organization of biomass of the species in both lateral spread and vertical composition (Lefsky et al., 1999). This is most noticeable between conifers and deciduous trees, which have very different shape, branch structure, and leafing. Interception differences between species vary on a seasonal level due to leaf loss in deciduous species as well (Helvey and Patric, 1965b). Common interception levels can be found within certain forest types, but even this varies geographically due to differences in climate (Crockford and Richardson, 2000). Spatial differences in interception create difficulty in reliable direct measurement.

Canopy interception (I, mm) is typically measured as the difference between gross precipitation (P, mm) and the portion of rainfall that enters the watershed, throughfall (T, mm) and stemflow (S, mm), and is represented by the following equation:

$$I = P - (T + S) + L \tag{3-1}$$

Interception losses from leaf litter (L, mm) are included here but most studies omit them due to difficulty in measurement, high variability, and small contribution (Helvey and Patric, 1965a). Interception is rarely measured directly and it is appropriate to consider the process in terms of the effective rainfall (R, mm), defined as the net precipitation entering the soil and contributing to the water budget (Swank, 1968).

$$R = P - I = T + S - L \tag{3-2}$$

Some studies omit stemflow measurement as well as litter interception due to insignificant levels and difficulty in measurement. Variability in the measurement of leaf litter interception and stemflow and the low contributions of both, on the order of 0 – 9% of gross rainfall for stemflow and 2-5% for leaf litter interception (Helvey and Patric, 1965b; Helvey, 1971; Swank et al., 1972) can have a cancelling effect since canopy interception is largely driven by canopy storage and evaporation.

The throughfall and stemflow portions of incoming rainfall are typically measured by random placement of multiple rain gages under the canopy for throughfall and tree collars for stemflow to account for and minimize the effects of spatial variability in these processes (Calder and Rosier, 1976; Crockford and Richardson, 2000). The lack of a standard practice for gage type, number of gages, or placement within the watershed in interception studies has led to results that are difficult to extrapolate outside of the specific conditions studied due to the sources for error and variability between canopy and climate at different locations (Helvey and Patric, 1965a). Rigorous statistical assessment of a study is necessary to determine adequate sampling procedures to ensure confidence and precision in water budget calculations that can be compared to results from different sites (Zarnoch et al., 2002). Adequate sampling may still produce inaccurate measures of throughfall though as values greater than gross precipitation have been reported (Crockford and Richardson, 2000). Many research efforts have focused on the development of a model that is able to accurately account for these physical and spatial variables and produce reliable estimates for canopy interception across all conditions due to the rigorous measurements needed to obtain acceptable accuracy. These models focus on the physical processes of canopy interception and vary greatly in simplicity, accuracy, and applicability.

The most widely used models of interception have been empirical regressions between interception (I) and gross precipitation (P) of the form:

$$I = aP + b \tag{3-3}$$

In this model, *a* and *b* are regression coefficients obtained through linear regression (Gash, 1979). Swank (1968) traces the development of these linear regression equations back to the work of Horton (1919), who quantified the physical process of interception loss (I, mm) as

$$I = S_j + K_1 E_r T_s \tag{3-4}$$

where S_j is interception storage capacity (mm), K_1 is the ratio of evaporative surface over the projectional area, E_r is the evaporation rate (mm/hr) during the storm, and T_s is the duration of the storm (hr). Swank (1968) goes on to describe the work of Kittredge (1948), who rewrote the equation to include the influence of gross precipitation (P, mm) as

$$I = S_j + (K_1 E_r T_s / P)P \tag{3-5}$$

This interpretation apportions interception into a canopy storage component and an evaporation component. If the evaporation component during any given storm is assumed to be a constant proportion of rainfall, interception becomes a linear function of gross precipitation represented by Equation 3-3. Substitution of Equation 3-3 for interception into Equation 3-2 for R (and assuming that S and L are negligible) produces a linear expression for throughfall (T) as a function of P.

$$R = T = P - I = P - (aP + b) = (1-a) * P - b \qquad (3\text{-}6)$$

This linear regression as an expression for T approximates the portion of rainfall not lost to evaporation (1-a) that is a function of P, decremented by a measure of canopy storage (b). This simplistic approach to modeling interception by linear regression of measured P and T has been applied to many forest stands in different areas in the eastern and southeastern United States alone (summarized in Table 3.1).

Table 3.1. Summary of throughfall regression equations for studies on tree species/ communities common to the lower coastal plain of South Carolina.

Study	Location	Vegetation Type	Throughfall Regression
Blood et al., 1986	Coastal South Carolina	Loblolly pine (*Pinus taeda*)	0.76P-0.06
Helvey, 1971	Multiple	Loblolly pine (*Pinus taeda*)	0.80P-0.01
	Mutiple	Average Pines	0.86P-0.04
Helvey and Patric, 1965b	Eastern United States	Hardwoods (Growing Season)	0.90P-0.03
		Hardwoods (Dormant Season)	0.91P-0.02
Roth and Chang, 1981	East Texas	Longleaf pine (*Pinus palustris*)	0.90P+0.10
Swank et al., 1972	Piedmont South Carolina	Loblolly pine (*Pinus taeda*), 5 yr.	0.83P-0.03
		Loblolly pine (*Pinus taeda*), 10 yr.	0.73P+0.00
		Loblolly pine (*Pinus taeda*), 20 yr.	0.76P+0.01
		Loblolly pine (*Pinus taeda*), 30 yr.	0.85P+0.00
		Hardwood/pine mix, mature	0.87P-0.02

These studies characterize the difference in water budgets with respect to canopy interception for stands of different forest communities, species type, age within species type, silvicultural treatment, and land cover change. All of them have shown variability in regression equations that are attributed to differences in stand properties and the influence of physical structure on canopy interception.

OBJECTIVES

The goal of this study is to assess the potential for spatial variability in rainfall measurement and throughfall on the UDC watershed to ensure rainfall measurements are accurate and determine if the canopy interception at UDC requires increased monitoring efforts for studying the rainfall response. Specific research objectives were to

1. Assess spatial variation in open-field measurements of event precipitation to determine if a single gage is adequate for rainfall measurements at UDC.
2. Calculate linear regressions for measured throughfall as a function of storm event rainfall for 10 subcanopy rain gages and determine if there is a seasonal difference in measured throughfall.
3. Compute an estimate for expected canopy interception at UDC using previous studies of throughfall on similar vegetation composition and forest communities.

METHODS

Site Description

Upper Debidue Creek (UDC) in Bannockburn Plantation (33.38° N, 79.17° W), located in coastal Georgetown County, is a 100 ha freshwater non-tidal watershed that has been slated for development. The location of UDC on the South Carolina coast and its proximity to development below the watershed outlet and downstream tidal marshes is shown in Figure 3.1.

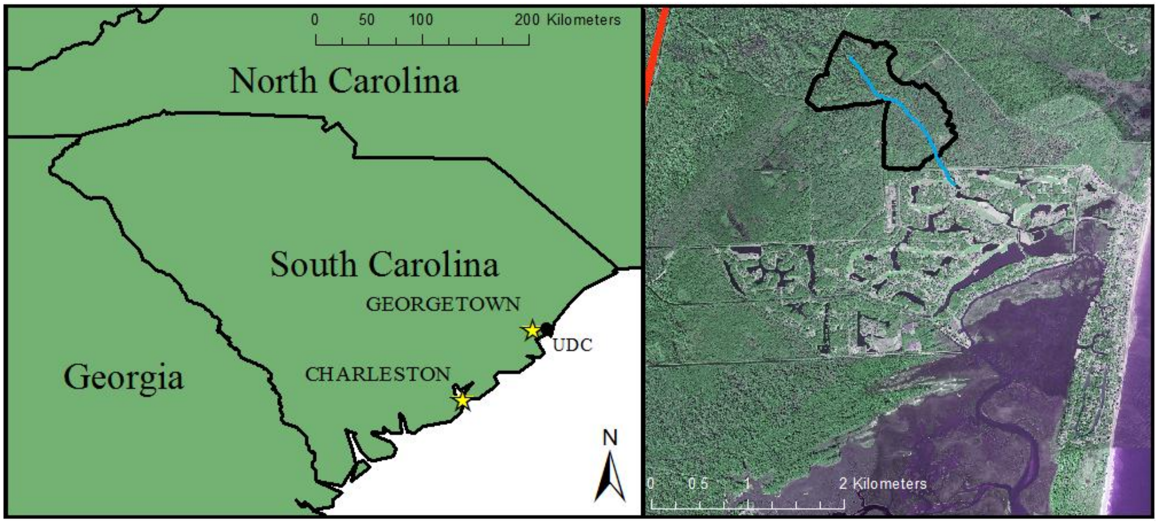

Figure 3.1. Upper Debidue Creek watershed, located in coastal South Carolina, is slated for development.

The landscape is currently dominated by forested wetlands with mixed hardwood lowlands and upland pine stands. The watershed is located in a humid subtropical climatic zone characterized by short mild winters and long hot summers and is prone to recurrent tropical storms during the fall months (La Torre Torres et al., 2011). The growing season is defined between calendar dates calculated as having a 50% probability to record the last frost of winter (sub 0° C) and first frost of fall from long-term local temperature records. These dates are from March 11 to November 20 at Georgetown, SC for UDC (NCDC, 1988). Dates outside this range represent the dormant season.

Data Collection

Rainfall and throughfall were measured for one year from 2011-2012. Tipping bucket rain gages (Onset® Hobo™, Bourne, MA at UDC Hobo and Middleton Creek locations, Rainwise® Rain Collector with Data Loggers™, Bar Harbor, ME at all other locations) located in four open field locations and ten subcanopy locations were used to quantify local hourly rainfall totals (Fig. 3.2).

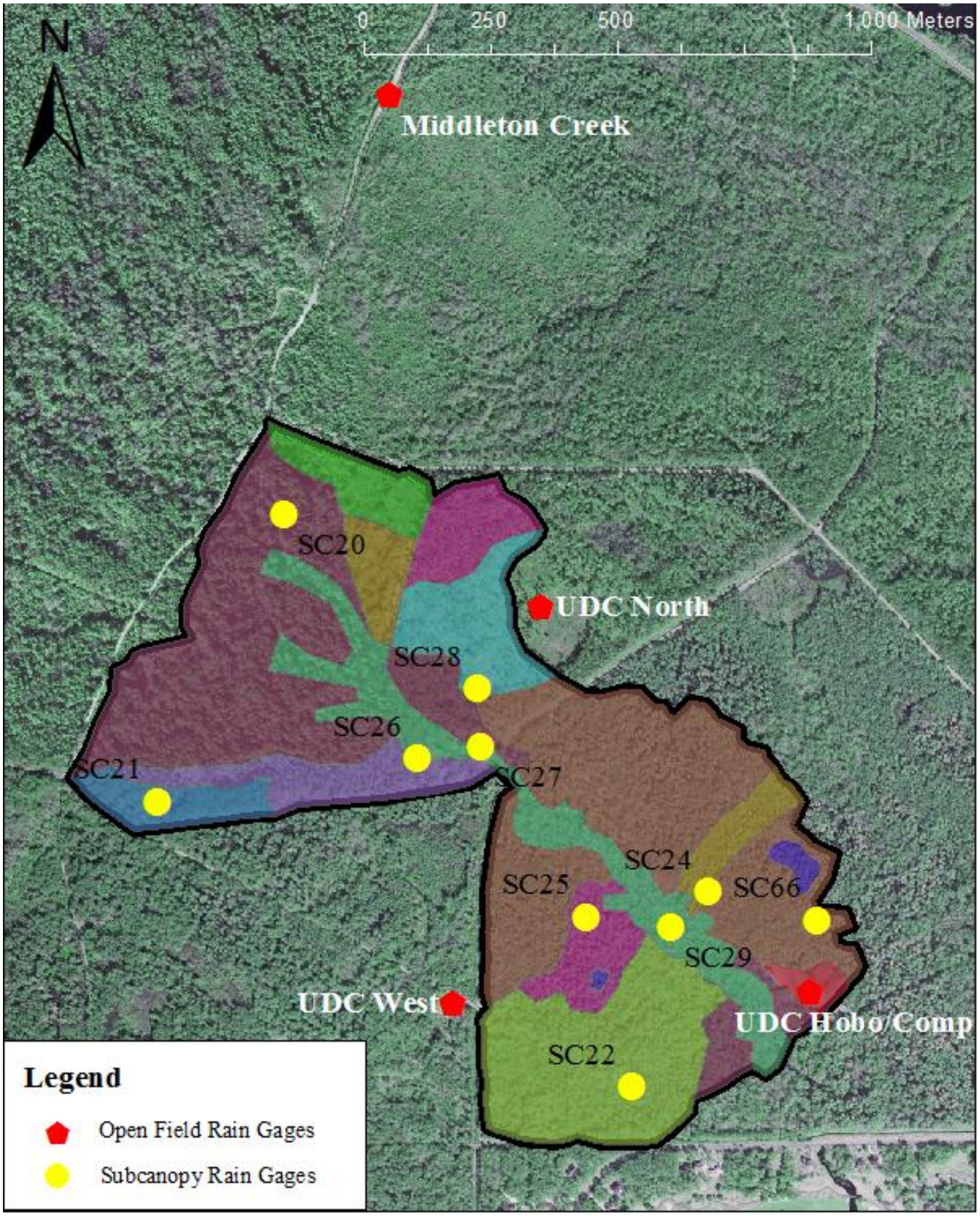

Figure 3.2. Upper Debidue Creek rain gage network with different vegetation communities. See Table 3.4 for descriptions of these vegetation communities.

The Middleton Creek and UDC Hobo rain gages have been monitored prior to this study. UDC Comp was installed on February 25, 2011 within 2 m of UDC Hobo. It serves as a comparison to UDC Hobo which has been used as the source for rainfall measurements at UDC in previous studies. The UDC West and UDC North rain gages were installed on March 22, 2011 in open field locations surrounding the UDC watershed in order to assess spatial variability in rainfall. The three gages were installed to a height of approximately 1.8 m with 45° sight lines that did not intercept the surrounding canopy. UDC West had to be removed for two periods of 4 days in April and 18 days in May due to controlled burning in the area. All rain gages were monitored on a monthly basis to ensure proper logging and to maintain gages so that clogging was avoided. Monthly datasets for rain gages that experienced clogging were eliminated from analysis.

Data Assessment

Hourly rainfall totals over the year for each rain gage were analyzed for comparison between the UDC Hobo gage and all other locations. The open field rain gages were each compared separately to the UDC Hobo gage in order to determine if there was a measurable difference between rainfalls measured at each location that would warrant the inclusion of a spatial network of rainfall measurements for further studies at UDC. Data was filtered to include only hours that at least one of the rain gages measured rainfall. The difference in rainfall measured between the gages for each of these hours was tabulated and the mean difference (μ_d) was analyzed. Measurement accuracy for the gages is ±1.0% for up to 1 in/hr for the Hobo® gages and ±1.5% for up to 0.5 in/hr for the Rainwise® gages. This represents a measurement error of ±0.01 in. for any given hour with measurement less than the stated intensity limits. Rain gage accuracies and tip increments were calibrated using U.S. customary units and all rainfall analysis and reporting will maintain these units. Hourly rain data was filtered again to include only hours that rainfall measured less than 1.0 in. by the Hobo® gages and less than 0.5 in. for the Rainwise® gages so that measurement accuracy was ensured. Evaluation of μ_d was performed using a two-tailed t-test to determine if it was non-zero (H_0: $\mu_d = 0$, H_a: $\mu_d \neq 0$). The t-statistic was calculated as

$$t = \frac{\mu d - D}{s/\sqrt{n}} \tag{3-7}$$

D represents the difference in the mean that is tested (0 in this case), s is the standard deviation of the sample, and n is the number of observations in the sample. The difference tested (D) was equal to 0. The null was rejected if the t-statistic was greater than 1.96 which represents the two-sided ($\alpha = 0.025$) significance level for a t-test with high degrees of freedom (n-1). A one-tailed t-test was used to determine if μ_d was less than 0.01 in. (H0: $\mu_d \leq 0.01$, Ha: $\mu_d > 0.01$). The difference being tested was 0.01, and the null hypothesis was rejected if the $t \geq 1.645$. This t-value represents the positive tail (α=0.05) significance level that indicates that μ_d is greater than 0.01 for a t-test with high degrees of freedom. The 0.01 in. threshold, equivalent to one tip for these rain gages, was chosen because it is the noted manufacturer's accuracy limit for each of the rain gages. Additionally, hourly rainfall data for these gages often shows a single tip on one gage that is not recorded by any others for a given hour. These errant tips may be caused by wind, disturbance by wildlife, or moisture collection in the gage. A μ_d greater than 0.01 would signify that differences in precipitation measurement between two gages are above manufacturer's measurement accuracies and possible errant tips and would warrant further investigation for potential spatial differences in rainfall measurement.

Hourly rainfall totals for the UDC Hobo rain gage and the ten subcanopy rain gages were processed to provide storm event totals. Storm events were bracketed by identifying periods of rainfall that were separated by at least 6 hours on each side with no rainfall. This approach is consistent with storm bracketing performed by Swank et al. (1972). Hours with measured rainfall of only 0.01 in. were not considered in determining the storm event window so that only the major portion of rainfall for the storm event was bracketed. Gross precipitation was measured at the UDC Hobo gage and storm events for this open field gage were compared to storm event rainfall for each of the subcanopy gages separately. Linear regression was performed to model storm event throughfall at each subcanopy gage as a function of gross precipitation. To assess seasonal differences in throughfall, storm events were separated into dormant season and growing season storms according to date. Linear regression was performed separately for each season to determine if there was a difference measured at each subcanopy gage.

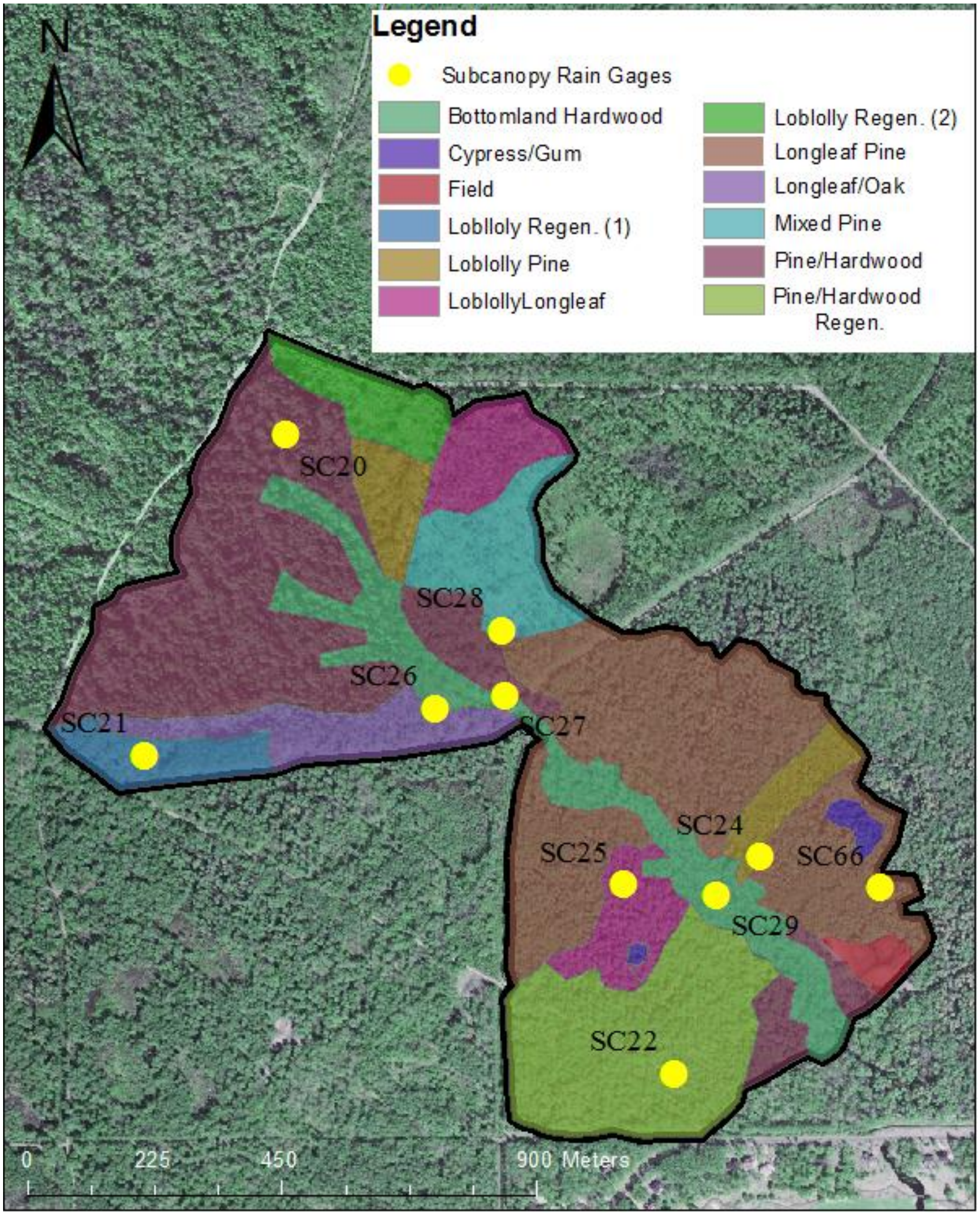

Figure 3.3. Areal vegetation composition of the Upper Debidue Creek watershed by dominant community.

The vegetation composition at UDC was previously assessed and appears in Figure 3.3 (provided by Dr. Williams from work with Bo Song, Clemson University, Belle W. Baruch Institute of Coastal Ecology and Forest Science, received August, 2011). Published values for expected throughfall for similar stands were compiled for each of the different communities represented on UDC (Table 3.1). Areas for each plant community were assessed using GIS to determine the area that each covered on the watershed. These areas, as a percentage of the whole watershed, were used with literature throughfall regression equations to calculate a single composite throughfall regression for UDC.

RESULTS AND DISCUSSION

The mean difference between hourly rainfall measured at UDC Comp, Middleton Creek, UDC North, and UDC West and the UDC Hobo gage was not higher than 0.01 in. for storm events during 2011-2012 (Table 3.2).

Table 3.2. Summary of descriptive statistics related to the mean difference in rainfall totals (in.) by storm event between all open canopy rain gages and UDC Hobo gage. (FTR indicates that statistical analysis failed to reject the null hypothesis).

Statistic	UDC Comp	Middleton Creek	UDC North	UDC West
Mean Difference (μ_d)	0.003	0.006	0.002	0.008
Standard Error	0.001	0.003	0.002	0.002
Standard Deviation	0.014	0.072	0.049	0.030
Minimum	-0.07	-0.43	-0.64	-0.15
Maximum	0.07	0.71	0.19	0.16
Sum	1.35	2.76	0.83	2.54
Count (n)	388	456	392	312
t(observed, H_0: mean = 0)	4.93	1.79	0.85	4.81
p-value	$p < 0.01$	FTR	FTR	$p < 0.01$
t(observed, H_0: mean $\leq$ 0.01)	-9.24	-1.17	-3.16	-1.10
p-value	FTR	FTR	FTR	FTR
Higher Gage for μ_d	UDC Comp	UDC Hobo	UDC North	UDC West

There was sufficient evidence that μ_d for the UDC Hobo and UDC Comp gages was non-zero ($p < 0.01$) but less than 0.01 ($p < 0.01$). The two closely placed gages appear to measure similar rainfall totals on an hourly basis. These results verify previous measurements made by the UDC Hobo gage and support the continued use as the main recording gage for the UDC watershed. Comparisons between the UDC Hobo gage and the Middleton Creek gage signify that these gages measure similar hourly rainfall totals on average. There was insufficient evidence to conclude that μ_d for these two gages was non-zero or greater than 0.01. There is the potential for spatial differences between these two gages which lie approximately 2 km apart. The range in difference between gage measurements for all storm events was wider than for other gages (-0.43 to 0.71). These larger differences appeared to be associated with dissimilar storm events measured at the two gages when storm event records were investigated. The location of the Middleton Creek rain gage (Fig. 3.2) is approximately 0.75 km north of the watershed. Despite this distance, the gages appear to measure similar hourly rainfall over the entire dataset. The UDC North rain gage lies between the Middleton Creek and UDC Hobo gages and is in close proximity to the northern areas of the watershed. There was insufficient evidence to conclude that μ_d was not equal to zero or greater than 0.01 in. Statistical evidence of the low μ_d for UDC Hobo and UDC North demonstrates similar rainfall measurements on an hourly basis between the two. A Thiessen distribution was generated for the four open field rain gages using ArcGIS 10. This distribution demonstrates the areas of the watershed that are best accounted for by each of the rain gages if they were all to be used for rainfall measurement. This distribution indicates that the area that the Middleton Creek gage would best represent in conjunction with the other three gages lies entirely outside of the UDC watershed boundary (Fig. 3.4). It is likely that the UDC North rain gage captures rainfall measurements for the northern end of the watershed more closely than the Middleton Creek gage because of the closer proximity demonstrated by the Thiessen distribution.

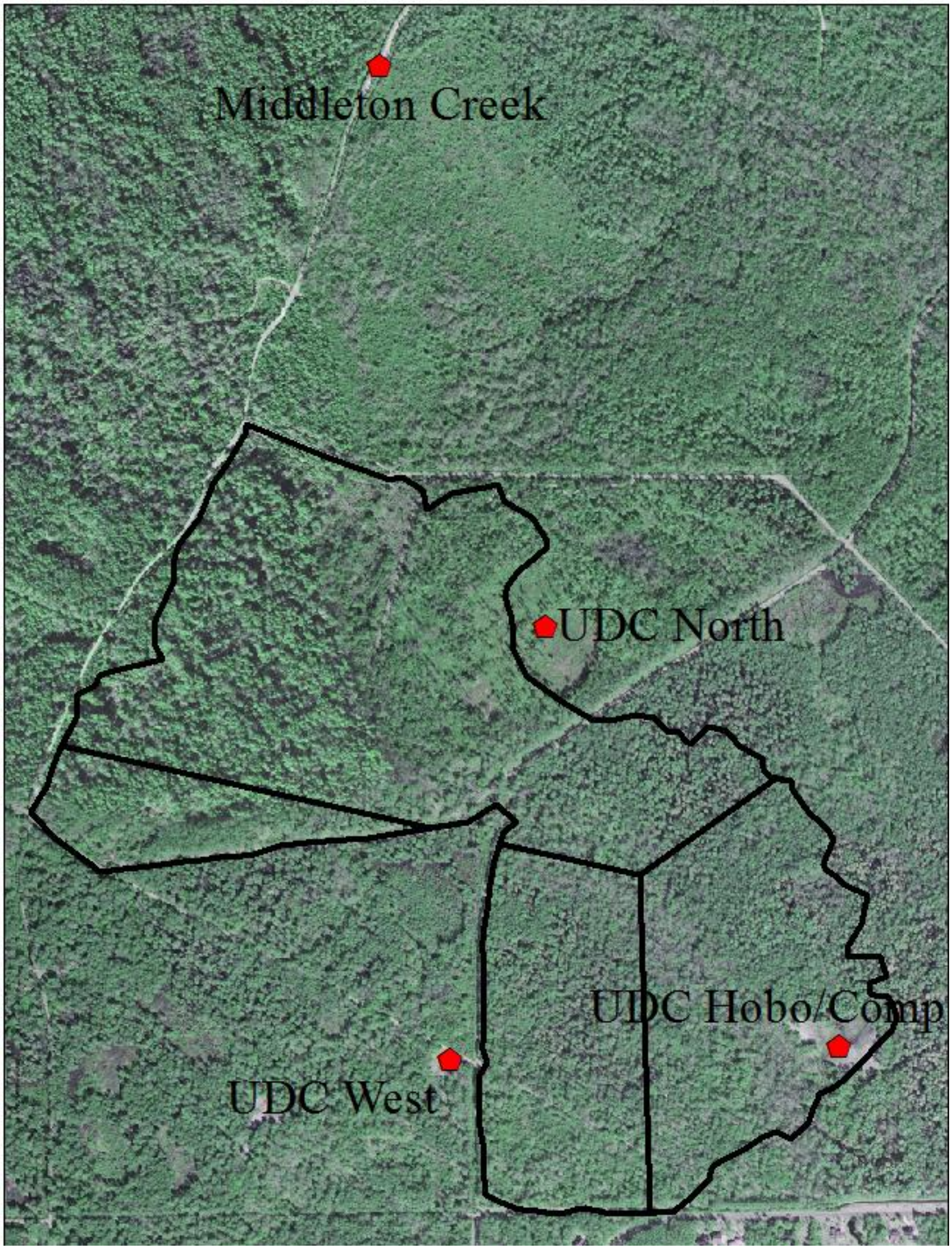

Figure 3.4. Thiessen distribution clipped to the Upper Debidue Creek watershed does not include any coverage by the Middleton Creek rain gage.

Comparisons between the UDC Hobo and UDC West gages indicate that hourly rainfall measurements at UDC West may be slightly greater than zero. There was sufficient evidence to conclude that μ_d was non-zero ($p < 0.01$) but insufficient evidence that μ_d was greater than 0.01 in. The mean difference was 0.008 in. and the range in measured differences in hourly rainfall was -0.15 in. to 0.16 in. The UDC Hobo gage is located in a much larger clearing, and the effect of wind may impact rainfall measurements for similar storm events between these locations. Also, the mean rainfall measurements of all three Rainwise® rain gages (UDC Comp, UDC North, and UDC West) were higher than the two Hobo® rain gages (UDC Hobo and Middleton Creek). For each gage comparison, the gage with the lower average mean was subtracted from the gage with the higher average mean in order to maintain positive results for μ_d. This difference is not by itself conclusive but there is the potential for systematic differences in measurement because of the different gages and their measurement accuracy. It does not seem that differences measured between the UDC Hobo and the other open field rain gages are sufficient to require additional rainfall measurements at UDC. The UDC Hobo gage measures rainfall for the watershed close enough to the different gage locations around the watershed that it alone can stand for storm event rainfall at UDC.

Subcanopy Gages and Throughfall

Throughfall regressions were compiled for each subcanopy gage over 2011-2012 storm events and separately for storm events during the dormant and growing seasons. They have been summarized in Table 3.3. They indicate a range of 0.76 to 1.24 for the slope parameter for the subcanopy gages over all storm events. This range is indicative of results that did not account for the spatial variability of throughfall. A slope parameter of greater than 1.0 indicates that the subcanopy gage measured greater rainfall totals than open-field rainfall measurement. Slope parameters were higher for dormant season storms at all subcanopy gages, and this indicates a seasonal difference in measured throughfall that may be related to physical canopy characteristics or climate variability.

Table 3.3. Summary of linear regression results of throughfall as a function of gross storm event precipitation for subcanopy gages at the Upper Debidue Creek watershed. P-values for the R^2 and slope parameter have been omitted for redundancy because they measured less than 0.01 for all instances.

Rain Gage	Season	R^2	Slope	Intercept	Intercept p-value
SC20	Overall	0.92	0.77	0.03	0.23
	Dormant	0.97	0.83	0.02	0.18
	Growing	0.92	0.77	0.03	0.39
SC21	Overall	0.91	0.98	0.03	0.27
	Dormant	0.98	1.05	0.00	0.95
	Growing	0.91	0.97	0.05	0.28
SC22	Overall	0.98	1.03	-0.03	0.03
	Dormant	0.99	1.13	-0.04	$p < 0.01$
	Growing	0.98	1.02	-0.04	0.07
SC24	Overall	0.96	1.05	-0.06	0.01
	Dormant	0.91	1.31	-0.10	0.02
	Growing	0.98	1.03	-0.07	$p < 0.01$
SC25	Overall	0.98	1.02	-0.01	0.48
	Dormant	0.97	1.15	-0.04	0.08
	Growing	0.99	1.01	-0.01	0.68
SC26	Overall	0.94	1.00	-0.03	0.19
	Dormant	0.93	1.33	-0.07	0.04
	Growing	0.95	0.99	-0.04	0.14
SC27	Overall	0.93	0.93	-0.01	0.71
	Dormant	0.95	1.18	-0.05	0.07
	Growing	0.94	0.88	0.00	0.88
SC28	Overall	0.93	0.76	0.03	0.21
	Dormant	0.95	1.00	-0.02	0.45
	Growing	0.93	0.74	0.02	0.37
SC29	Overall	0.97	1.24	-0.12	$p < 0.01$
	Dormant	0.91	1.42	-0.12	0.02
	Growing	0.98	1.24	-0.13	$p < 0.01$
SC66	Overall	0.98	0.96	0.00	0.92
	Dormant	0.99	0.99	-0.01	0.51
	Growing	0.98	0.96	0.00	0.85

The p-values have been omitted for the R^2 and slope parameter because they were less than 0.01 for all instances. These results indicate that the regressions demonstrate a significant relationship between storm event precipitation and throughfall measured. All regressions had a coefficient of determination greater than 0.90 and account for a large amount of variation in throughfall measured. P-values for intercept variables varied largely and were mostly greater than 0.05, so the significance of this regression parameter was not high for most regressions. P-values for the slope parameter were all less than 0.01, indicating the significance of these parameter estimates. The significance of this parameter is important because it approximates the percentage of gross rainfall measurements that is measured as throughfall. Results indicate that several of the subcanopy gages recorded higher throughfall than gross precipitation. The higher readings were likely caused by significant drip from the canopy after canopy saturation has occurred. Static locations used for throughfall measurement are subject to potential drip that is likely to occur due to the position of the gage under the canopy. Therefore, results from static point measurements may not be indicative of the areal distribution of throughfall. Location of each of the subcanopy gages in relation to vegetation community can be seen in Figure 3.3.

Five of the subcanopy gages demonstrated throughfall less than gross precipitation over all storm events (SC20, SC21, SC27, SC28, and SC66). A sixth gage, SC26, indicated that throughfall was equal to gross precipitation decremented by a constant 0.03 in., though this intercept parameter was not significant ($p = 0.19$). It is possible that this may represent a canopy storage capacity for that specific location that is filled before throughfall occurs. The SC20 rain gage measured a throughfall regression with a slope parameter of 0.77. This gage is located in a dense pine/hardwood stand and this measurement seems reasonable. Swank et al. (1972) measured a throughfall regression of 0.87P-0.02 for a pine/hardwood stand dominated by oaks in the piedmont of South Carolina. Differences in throughfall measured may be attributable to stand composition and climate differences between the two sites. The throughfall regression for SC21 measured less than gross precipitation overall with a slope parameter of 0.98. This gage was located in a stand of immature loblolly pine. Swank et al. (1972) measured throughfall on the order of 0.73P for a stand of 10-year old loblolly in the piedmont of South Carolina. It is likely that SC21 was subject to canopy drip at its static location. The SC27 gage is located in an area of riparian hardwoods and throughfall regression measured an overall slope parameter of 0.93. Helvey and Patric (1965b) measured average throughfall for all eastern hardwood species and found an average of 0.90P-0.03 for the growing season and 0.91P-0.02 for the dormant season. Results for SC27 show similarity to these measures over all storm events, but there is a large deviation for the dormant season regression with a slope parameter of greater than 1.0. The SC28 gage is located near SC27 in an area of mixed pines and measured the lowest throughfall regression over all storm events of all gages with a slope parameter of 0.76. Throughfall for this gage was lower than the average throughfall measured for pine species by Helvey (1971) of 0.86P-0.04. This gage is located in a stand that is not very dense, and open conditions may effect gage measurements due to greater wind influence. The SC66 gage is located in a stand of longleaf pine and measured a throughfall regression slope parameter of 0.96. This is slightly higher than measurements made by Roth and Chang (1981) who measured throughfall in a longleaf pine stand in east Texas of 0.90P+0.1.

Throughfall regression equations calculated for the ten static subcanopy gages at UDC display a range of canopy throughfall that changes between locations. This variability in throughfall measurements is likely related to the different canopy densities and vegetative communities at each subcanopy rain gage. These measurements are subject to drip because of the static location of the gages, though this was not observed for all gage measurements. Half of the gages had measured throughfall regression equations that indicated throughfall less than gross precipitation. These measurements were comparable to available literature studies on throughfall regressions for similar canopy composition though results are difficult to compare because spatial differences were not accounted for. It is difficult to make conclusions concerning this data but it is worth noting that for all subcanopy gages, throughfall regressions had smaller calculated slope parameters for growing season events (Table 3.3). This results indicates that the amount of throughfall calculated for dormant season storm events by regression of event data is a higher percentage of gross precipitation than for growing season storms. This is indicative of the role of greater vegetative surfaces and ET during the growing season and the effect that it has on throughfall levels. Results therefore demonstrate that canopy interception is higher during the growing season, as expected, and may play a role in intermittent streamflow dynamics that are observed in headwater catchments in the LCP.

Throughfall Estimates from Literature Regressions

Estimates for expected throughfall at UDC were calculated using throughfall regressions from previous studies (Table 3.1) and the vegetative composition of the UDC watershed (Fig. 3.3). Different plant communities were matched to literature equations of throughfall regressions for similar stands as closely as possible. The areal composition and throughfall regression equation used for each community is summarized in Table 3.4.

Table 3.4. Areal percentages and throughfall regressions used by vegetative community to determine expected throughfall for the UDC watershed based on previous studies in the southeastern United States.

Vegetation Type	Areal %	Regression	Source
Bottomland Hardwood	10.34%	0.90P-0.03 (GS); 0.91P-0.02 (DS)	Helvey and Patric, 1965b
Cypress/Gum	0.68%	0.90P+0.10	Same as longleaf pine
Open Field	1.03%	P	Open field precipitation
Loblolly Pine	2.20%	0.76P-0.06	Blood et al., 1986
Loblolly Pine Regen.	8.18%	0.73P+0.00	Swank et al., 1972; immature 10 yr.
Loblolly/Longleaf	6.32%	0.83P+0.02	Average of Longleaf and Loblolly
Longleaf Pine	23.71%	0.90P+0.10	Roth and Chang, 1981
Longleaf/Oak	4.66%	0.90P+0.10	Roth and Chang, 1981
Mixed Pine	5.14%	0.86P-0.04	Helvey, 1971; Average of pines
Pine/Hardwood	24.43%	0.87P-0.02	Swank et al., 1972; Pine/hardwood
Pine/Hardwood Regen.	13.30%	0.87P-0.02	Swank et al., 1972; Pine/hardwood
Composite	100%	0.87P+0.02	

No information for Cypress/Gum throughfall could be found. This community makes up less than 1% of the UDC watershed. The two stands that were observed by Williams and Song (2006) are surrounded by Longleaf pine stands, so the Longleaf pine regression was used for the Cypress/Gum throughfall for UDC. The different seasonal throughfall regressions noted by Helvey and Patric (1965b) for hardwoods were used to determine if a seasonal difference is to be expected at UDC. Results derived from the calculation of an area-weighted composite throughfall regression were the same for growing season and dormant season. These results estimate throughfall at UDC as 0.87P+0.02 for all storm events. This is in line with the 12% estimate of canopy interception as a portion of ET reported by Harder et al. (2007) on a similar forested headwater catchment in the LCP. It does not account for seasonal differences in throughfall that are observed when throughfall regressions were calculated separately for dormant and growing season storm events (Table 3.3). Seasonality was not included in any of the literature regression equations for throughfall except the bottomland hardwoods, and these represent only 10% of the UDC watershed. Regression equations used from other studies are typically only applicable to sites in close proximity to where they were calculated due to the large variability in canopy interception variables. A study performed on loblolly pine stands in the LCP yielded different throughfall regressions than were calculated for similar stands in the piedmont of South Carolina, but they were close enough to be considered comparable (Blood et al., 1986). Differences in climate may produce the greatest difference in canopy interception processes because of the role that rain patterns and evaporation play. The seasonal differences in measured throughfall regressions for UDC indicate that the presence of hardwoods may not be the only factor contributing to seasonal differences in throughfall. Trends in ET and a long growing season may also have an impact on the level of canopy interception on these watersheds. Seasonal differences should be considered along with other seasonal trends in LCP hydrology for the role that they may play in forested watershed streamflow dynamics.

CONCLUSIONS

Hourly measurements of rainfall by the UDC Hobo rain gage appear sufficient to characterize rainfall for the entire UDC watershed. Differences between this gage and other gages surrounding the watershed were either within measurement accuracy limits for the gages or not substantial enough to warrant a spatial network of rainfall measurements. Throughfall measurements at ten static subcanopy rain gage locations within the UDC watershed are not accurate enough to support conclusions on throughfall in this watershed due to the influence of canopy drip. The static locations do not accurately assess the spatial distribution of throughfall under the canopy. Using previous throughfall regression studies for comparable vegetative communities on UDC, an estimate for throughfall was obtained that is similar to measurements calculated on a comparable LCP watershed through water budget calculations. This regression may serve as an approximation for throughfall at UDC. Throughfall regressions for dormant and growing season storm events highlight a seasonal difference in measured throughfall at all gage locations. This seasonal difference in throughfall may relate to ET trends and the long growing season. This seasonal difference should be considered when examining trends in LCP forested headwater streamflow dynamics because it may influence water budgets and the storm response on these watersheds. Canopy interception that is higher during the growing season lessens the effective rainfall that enters the watershed, and this watershed loss from rainfall should be accounted for in water budgets.

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CHAPTER FOUR

CHARACTERIZATION OF STORM FLOW DYNAMICS OF HEADWATER STREAMS IN THE SOUTH CAROLINA LOWER COASTAL PLAIN

(Epps et al., in review)

ABSTRACT

In coastal watersheds, streamflow is influenced by fluctuating shallow groundwater levels which create high variability in outflow response to storm events. For effective coastal land use decision-making, baseline watershed hydrology must be assessed as a benchmark for stormwater management goals, especially with respect to seasonal and event-based groundwater influences on these outflows. Toward the goal of quantifying these water budgets and flow dynamics, hydrologic monitoring has been conducted in two first-order lower coastal plain watersheds located in the southeastern U.S., a region experiencing a tremendous amount of growth and development. Data from storm events over a three-year period were analyzed to determine direct runoff estimates (ROC) and the total storm outflow response (TSR) as a percentage of rainfall. ROC calculations were accomplished using empirical and graphical hydrograph separation techniques that partition total streamflow into sustained base flow and direct runoff components. Based on these outflow allocations, the ROC ratios ranged from 0 to 0.32 on the Upper Debidue Creek watershed (UDC) and from 0 to 0.57 on Watershed 80 (WS80). TSR calculations represent a greater portion of streamflows, and results display a much greater range than ROC for both watersheds. These measures ranged from 0 to 0.93 at UDC and 0.01 to 0.74 at WS80. Variability in event runoff generation is

attributed to seasonal trends in soil moisture conditions that are regulated by the balance between rainfall and evapotranspiration. Water table elevation was shown to influence streamflows in the study watersheds. By examining relationships between antecedent water table elevation, streamflow, the ROC, and TSR, break points within each watershed were identified that represent a threshold groundwater elevation above which event runoff generation and outflows increase sharply in response to rainfall. These relationships that influence runoff generation mechanisms are unique to the lower coastal plain and driven by variable groundwater elevations.

BACKGROUND

Coastal headwater streams in undeveloped forested landscapes function as a natural storage and conveyance mechanism for groundwater discharges and streamflow (Amatya et al., 2006). The Lower Coastal Plain (LCP) of South Carolina is defined by low gradient topography and low elevations typical of southeastern United States coastal landscapes. Shallow groundwater elevations influence soil moisture levels and couple with surface water generation during rainfall events to determine stream outflows that include significant base flows (BF) (Eshleman et al., 1994; Williams, 2007). The magnitude of watershed outflows is often driven by a fluctuating water table position that is regulated by the balance between evapotranspirative demand and infiltrative replenishment by rainfall (Miwa et al., 2003; Amatya et al., 2006; Slattery et al., 2006; Harder et al., 2007). High water table elevation and high soil moisture conditions lead to higher outflow production during winter months when forest vegetation is largely dormant and evapotranspiration rates are lower (Harder et al., 2007; Williams, 2007; Amatya and Skaggs, 2011 La Torre Torres et al., 2011). During summer months, streamflows are intermittent in response to direct rainfall. High summer evapotranspiration rates tend to rapidly lower the water table elevation resulting in increased soil storage and decreased storm runoff (Slattery et al., 2006; Harder et al., 2007; Williams, 2007; Amatya and Skaggs, 2011). Flow cessation occurs when the water table elevation is sufficiently low that groundwater flows are disconnected from the stream channel. Between these seasonal extremes, the rainfall response is dependent upon antecedent moisture conditions (AMC) that vary with microclimate variability and seasonal evapotranspiration shifts (Sun et al., 2002; Harder et al., 2007). Due to these highly variable conditions, the derivation of water budgets for coastal forested watersheds with low-gradient topographic relief can be complex.

It is critical to better understand these hydrologic dynamics for the protection of water resources and flood prevention in coastal landscapes, especially as forested areas are being converted to residential and commercial development. Developing lands are prone to increased impervious surface area which can significantly alter site hydrology (magnitude and pathways of surface and subsurface flow) and increase pollutant loads to adjacent waters (Arnold and Gibbons, 1996; Booth et al., 2002). A better understanding of baseline outflow production on these headwater streams is needed. Land use in the region is changing rapidly with growing populations and an intensive timber industry, and this change has motivated the study of pre-development conditions in the area to ensure water resource protection (Allen and Lu, 2003; Amatya and Skaggs, 2011; Blair et al., 2011). Increased timber harvesting and land cover changes associated with urbanization in the area will impact the quantity and timing of runoff as well as the water quality, and these can alter established patterns of flow (Harder et al., 2007; USFS, 2011). The establishment of a baseline for the hydrologic characteristics of coastal headwater streams will guide effective future land use decision making to sustain watershed health and ecosystem function. Additionally, downstream tidal marsh ecosystems that are sensitive to impaired water quality from upstream sources are also of concern as land use changes take place in upland watersheds (Holland et al., 2004).

Though much work has been done to characterize upland watersheds, more information is needed specific to the LCP with respect to regional hydrologic processes (Amatya et al., 2006). Most recently, Amatya et al. (2011) summarized the studies concentrated on eco-hydrologic processes, restoration, and effects of management practices on hydrology and water quality of the forests and forested wetlands of the Atlantic Coastal Plain. Previous studies of headwater catchments in the LCP have reported variable annual outflows as a percentage of rainfall. Amatya et al. (2006) measured total annual outflow depth as a percentage of rainfall over a long-term dataset covering 23 years on two first-order forested watersheds in the Francis Marion National Forest located in the LCP of South Carolina. Results ranged from 5% to 59% on a control watershed (WS80) and from 9% to 44% on a treatment watershed (WS77). Differences in total outflow between years are due to variations both in the temporal distribution of annual rainfall and the AMC at the time of rain. In coastal forest water budgets, the relationship between rainfall and outflow production is affected by soil moisture levels that are influenced by the shallow water table (Amatya and Skaggs, 2011). Harder et al. (2007) computed water budgets for WS80 over two consecutive years and measured a total outflow depth as a percentage of rainfall of 0.47 for 2003 and 0.08 for 2004. This range in outflows between years was partially due to differences in annual rainfall (1670 and 960 mm, respectively) and partially due to differences in AMC. Several large storms during 2004 resulted in only moderate to low outflows, and this was linked to lower water table elevations at the time of rainfall that characterize dry AMC as was also shown by Dai et al., (2011) in their modeling study. In another study, Sun et al. (2002) compared the hydrologic response of two flat LCP watersheds in North Carolina (NC) and Florida (FL) to a high gradient watershed with considerable topographic relief in the Appalachian region of North Carolina (UP) using long-term precipitation and flow data. The UP watershed demonstrated higher results (0.53) compared to the two lowland watersheds (0.30 for NC, 0.13 for FL). Climate variability was one factor for the difference, with average annual precipitation of 1730 mm for the UP watershed and 1520 mm and 1260 mm for the LCP watersheds in NC and FL, respectively. Outflow in the high-gradient watershed had consistent BF contributions and flowed constantly during the study period. Intermittent flow was observed on the LCP streams reflecting variable water table elevation and intermittent groundwater discharges. Variable AMC affects BF levels as these conditions change and stream outflows behave accordingly. These observations are consistent with Todd et al. (2006) who reported that hydrologic processes in wetland-dominated basins are inconsistent with some aspects of the variable source area concept of streamflow generation. Some parts of the basin may become decoupled from the basin outlet as summer progresses. Runoff from these portions of the watershed may be lost to evaporation and infiltration or held in surface storage even before reaching the outlet.

Outflows on these LCP headwater streams are not determined by rainfall alone because of variable AMC consistent with Todd et al (2006) who noted that the vertical water movement due to rainfall, ET, and deep seepage was more important than the lateral groundwater flux in explaining the wetland's hydrologic behavior. Harder et al. (2007) showed that the temporal distribution of rainfall as it relates to AMC has more influence on outflow production than just rainfall totals alone. Sun et al. (2002) analyzed total outflows related to isolated storm events on the LCP FL watershed. Storm events were selected to assess the difference in outflows generated by both small and large storms that fell on the watersheds for both dry and wet AMC. The magnitude of streamflow prior to rainfall, relative to typical outflows on the watershed, was used to differentiate between dry and wet AMC. Small storms in the range of 30 – 59 mm demonstrated lower storm outflow depths on the FL watershed that were consistent with lower annual outflows. Large storms in the range of 102 – 160 mm demonstrated lower storm event outflows at the FL watershed for dry AMC (0.08 as a ratio of event rainfall) and much higher storm outflow depth for wet AMC (0.58 as a ratio of event rainfall). The large increase in outflow production in the rainfall response from dry AMC to wet AMC at FL demonstrates the role of soil storage on outflow production on these LCP headwater streams. The dry AMC is associated with lower water table elevations and higher soil storage that is filled before runoff is generated. The wet AMC is characterized by high water table elevations with low soil storage, and these conditions generate runoff rapidly under saturated conditions (Amatya and Skaggs, 2011).

Storm event outflow production must be studied in order to determine the role that seasonal trends in evapotranspiration, water table elevation, and AMC have on LCP headwater streams in the rainfall response. The storm event rainfall response is typically referred to as runoff, and this term often has different meanings from study to study. The majority of previous studies conducted in the LCP have defined runoff as total stream outflow depth associated with a given storm event and usually express it as a percentage of rainfall. One of the objectives for this study is to differentiate between BFs associated with groundwater discharge and direct runoff that is related to surface water generation. In this study, total outflow depth measured as a percentage of rainfall will be referred to as the total storm response (TSR) in order to differentiate between different measures of storm event runoff. Direct runoff estimates, also expressed as a percentage of rainfall, will be referred to as the direct runoff coefficient (ROC). A study by La Torre Torres et al. (2011) demonstrated that the TSR changes seasonally in step with trends in AMC. Storms selected from a long-term dataset at Watershed 78 in the Francis Marion National Forest in the LCP of South Carolina were separated according to the wet (Dec. – May) and dry seasons (June – Nov.). Results demonstrated a significant relationship between rainfall and TSR during the wet season ($r^2 = 0.68$, $p < 0.01$) and less so during the dry season ($r^2 = 0.19$, $p = 0.02$). Rainfall accounts for a lower amount of variability in TSR during the dry months due to lower water table elevation and higher soil storage caused by increased ET demands. The authors also suggested that the event runoff was controlled mainly by rainfall amount and the AMC represented by the initial flow rate. In a coastal North Carolina study, Amatya et al. (2000) found the drainage outflow as a percentage of rainfall varying between 2% for a dry summer period with water table as deep as 2.3 m to as much as 89% for a wet winter event with near surface water table for a drained pine forested watershed. These authors also reported that the amount of rainfall needed to generate a drainage event depends upon the AMC, which was represented by initial flow rate in their study. Williams (2007) observed a non-linear relationship between rainfall and direct runoff estimates in an LCP headwater stream. Runoff generation mechanisms unique to LCP hydrology were related to water table elevation. Base-flow levels were modeled by a graphical hydrograph separation method that accounts for accelerated groundwater discharges in the rainfall response. These flows were found to track groundwater elevations closely due to the relationship between water table elevation and watershed outflows for these streams as shown by Harder et al. (2007). Another investigation of the rainfall response on a LCP headwater catchment at Upper Debidue Creek (UDC) near Georgetown, SC showed that groundwater elevations tracked the stream outflow hydrograph closely and the ROC increased with consecutive closely spaced storms that contributed to wet watershed conditions (Rogers et al., 2009). The assessment of the rainfall response is complicated by the interaction between groundwater discharges and surface water generation. Water table dynamics that are influenced by climate and evapotranspiration contribute to variable AMC on LCP headwater streams. This results in a range of outflows as a percentage of rainfall on an annual basis as well as between storm events. Most recently, Callahan et al. (2011)

estimated total recharge to groundwater for the same LCP watershed studied by La Torre Torres et al. (2011) by analyzing water table response to storm events and the rate at which water was transferred into the shallow aquifer. The authors attributed the difference found in two methods of estimating the recharge (water table fluctuation method and Darcy equation) to the ET and AMC not accounted for in the Darcy method.

OBJECTIVES

The objective of this study is to assess the rainfall response on two headwater streams in the LCP using TSR and ROC as measurements of storm event runoff. The relationship between runoff measurements (TSR and ROC) and estimates of AMC using various methods outlined below will be determined to characterize differences in runoff generation related to seasonal trends and variable AMC from storm to storm. The role of groundwater in runoff generation will be primarily assessed in terms of water table elevation, and results will be used to assess runoff generation mechanisms in LCP headwater streams.

METHODS

Site Descriptions

The study sites are located at two first-order Lower Coastal Plain (LCP) watersheds. Upper Debidue Creek (UDC) in Bannockburn Plantation (33.38° N, 79.17° W), located in coastal Georgetown County, is a 100 ha freshwater non-tidal watershed that has been slated for development. Potential runoff created by development is of concern on this tract. Existing downstream development is already experiencing water quantity and quality issues that have forced overflow routing to nearby Winyah Bay to avoid discharges to protected downstream tidal marshes at North Inlet. Watershed 80 (WS80), a tributary of Huger Creek but adjacent to Turkey Creek located in the Francis Marion National Forest (33.15°N 79.8°W), is a 163 ha freshwater non-tidal watershed that is federally protected and serves as an undeveloped reference watershed. The location and monitoring design for each can be seen in Figure 4.1.

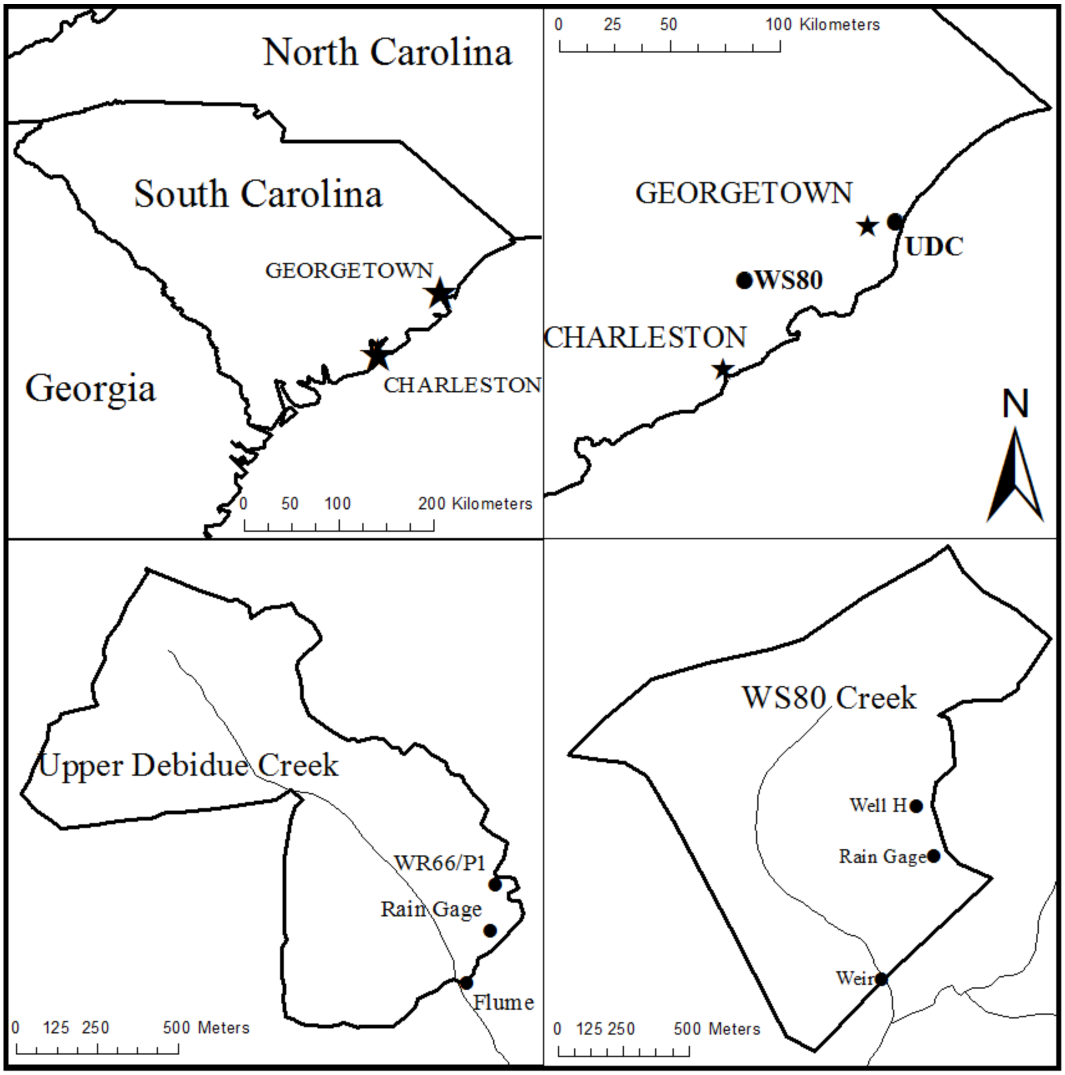

Figure 4.1. Upper Debidue Creek and Watershed 80 locations and monitoring networks.

Both watersheds are characterized by low gradient topography and shallow water table conditions. Surface elevations at UDC range from 6.5 m above sea level (asl) in the upland area to 2.1 m asl at the watershed outlet. Surface elevations on WS80 range from 10 m above mean sea level in upland areas to 3.7 m at the watershed outlet (Harder et al., 2007). The landscape is currently dominated by forested wetlands with mixed hardwood lowlands and upland pine stands. The primary soils in the UDC watershed are Lynn

Haven and Leon. These soils are formed of sandy marine sediment, are associated with very low gradient conditions, are highly permeable, and poorly drained (USDA, 1980). The primary soils in WS80 are Wahee, Meggett, Craven, and Bethera. These soils are formed of clayey Coastal Plain sediments and are typical of areas with low gradient topography (USDA, 1974). These soils are poorly drained with high available water content and have lower permeability than sandy soils. The two watersheds are about 75 km apart and in a humid subtropical climatic zone characterized by short mild winters and long hot summers (La Torre Torres et al., 2011). The growing season is defined between calendar dates calculated as having a 50% probability to record the last frost of winter (sub 0° C) and first frost of fall from long-term local temperature records. These dates are from March 9 to November 25 at Charleston, SC for WS80 and from March 11 to November 20 at Georgetown, SC for UDC (NCDC, 1988). Dates outside this range represent the dormant season.

Data Collection

Rainfall, streamflow, and groundwater elevation data from the two study watersheds were collected from 2008-2011 (Fig. 4.1). Tipping bucket rain gages (Onset® Hobo™, Bourne, MA at UDC and WS80) located in both watersheds were used to quantify local hourly rainfall totals. Groundwater elevations were monitored at upland locations near the watershed boundary on both UDC and WS80. At UDC, a 3 m deep water table well (4.2 m asl) with a pressure transducer was located in an upland pine area near the watershed boundary (WR66, Fig. 4.1). The pressure transducer was replaced with a Solinst logger in March 2011. Groundwater elevation was monitored in the upland area at WS80 (Well H, Fig. 4.1, 9.09 m asl) by a WL16 logger (Global Water, Gold River, CA). Watershed outflow in UDC was estimated using a 0.6 m modified Parshall flume located immediately downstream of a road culvert. Additional instrumentation details for UDC are provided by Hitchcock et al. (2009). Streamflows were corrected for submergence in the flume according to equations developed by Peck (1998). A threshold of 0.85 for submergence was set for measured outflow in accordance with the correction equations. At WS80, streamflow rate was estimated by measuring stage over a compound weir. Additional instrumentation details for flow measurement in WS80 are provided by Harder et al. (2007). Storm-flow rates were first converted to volumes by hydrograph integration. They were converted to equivalent depths in millimeters by dividing the runoff volume by watershed area.

Data Assessment

ROC and TSR ratios were calculated using a graphical hydrograph separation method to determine the total storm-flow depth and to distinguish the amount of direct runoff from BF for a given rainfall event (Fig. 4.2a). The rationale for using this procedure was to calculate only the amount of flow generated as storm response quick flow, or direct runoff. The method differentiates direct runoff from the more delayed groundwater derived BF component of the TSR to obtain estimates of ROC. These two components of outflow discharge at different timescales as a signature of their generation mechanism in the rainfall response. Sustained BF contributions for LCP headwater streams are characteristic of groundwater elevation and watershed drainage and they may be influenced by previous storm events and AMC. ROC and TSR estimates will be compared to determine differences between the methods in regards to BF. Storm events were selected from streamflow data for 2008-2011 at both watersheds in order to assess the rainfall response over comparative local climate and moisture conditions because of their proximity. Storms were selected given that they met the following criteria: (1) rainfall greater than 20 mm, (2) outflow was generated, and (3) event displayed a single-peaked hydrograph with a recession limb of sufficient length to perform graphical analysis. These criteria allowed for selection of a sufficient number of storms from the three-year period at both watersheds that ranged from substantial, higher frequency storm events up to larger storms of much less frequency. The storm event at WS80 on February 2, 2011 displayed a secondary local peak during the recession limb that was not as high as the main peak in outflow. The additional rainfall occurred during saturated conditions and this additional peak was measured as DR for the storm event which is consistent with method assumptions. The AMC features measured for each storm event by compiling the initial outflow rate just prior to rainfall, the antecedent precipitation index (API) for 5- and 30-days, and the initial water table elevation prior to rainfall. API was calculated by summing rainfall amounts for 5 and 30 days prior to the storm event.

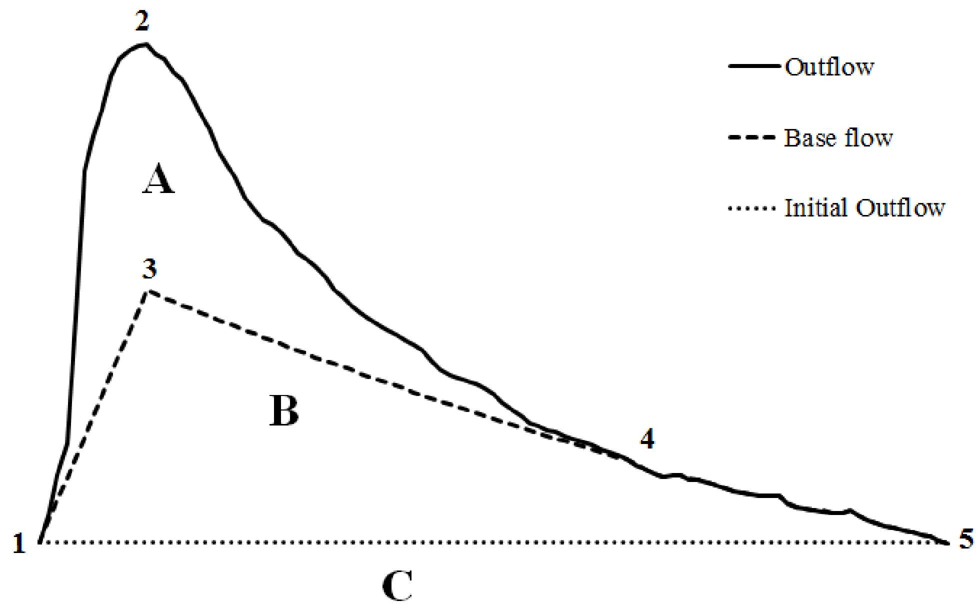

Figure 4.2a. Graphical representation of the hydrograph separation method for determination of ROC and TSR.

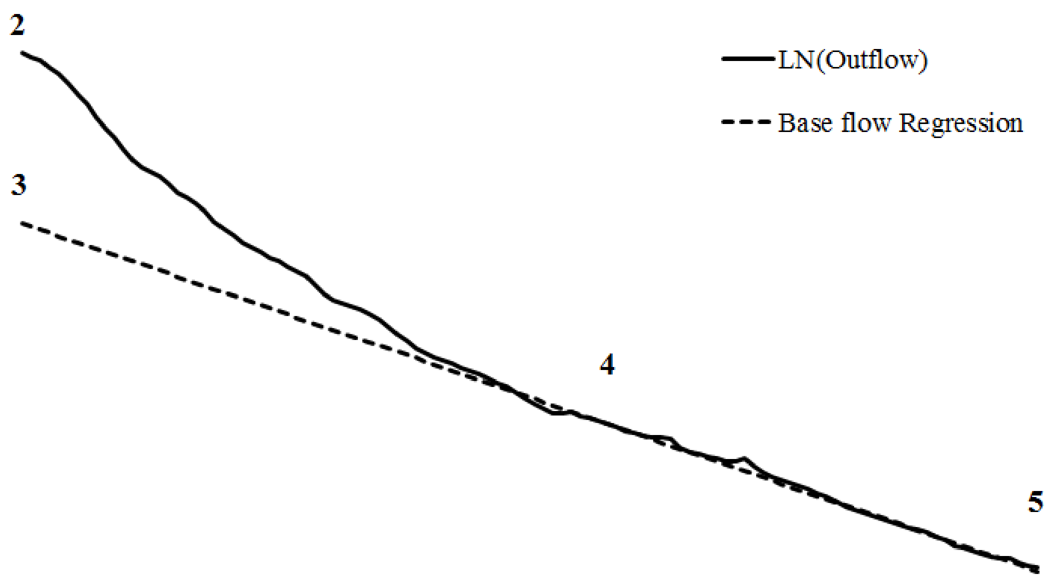

Figure 4.2b. Recession limb analysis for base flow separation in log scale. Number identifiers are equivalent to representative points in Fig. 4.2a.

Hydrograph separation was performed based on a method developed by Williams (2007) that emphasizes the physical processes of lower coastal hydrology, and it is represented graphically in Fig. 4.2a. This method models BF contributions to outflow that increase with rising water table elevation (Fig. 4.2a, segment 1-3) in response to rainfall. BF increases up until the peak outflow (Fig. 4.2a, point 2) is reached. Peak BF is followed by a sustained recession related to typical groundwater discharges (Fig. 4.2a, segment 3-5) that result from gravity-driven watershed drainage as the water table lowers after rain ceases. This method models total outflows as the sum of two parallel linear reservoirs that discharge at different rates. A longer, slower outflow is expected for groundwater inputs and is considered here as BF. BF volumes are represented in Fig. 4.2a as the sum of areas B and C. The faster component, termed "quickflow", is taken as a measure of direct runoff (DR). DR is represented by area A in Fig. 4.2a. Williams uses the linear model of Maillet (1905) to model BF recession. This model describes outflow rate at a particular time as a function of initial flow rate, time and a rate constant. The model is expressed as:

$$Qt = Q0 * \exp(-t/k) \qquad (4\text{-}1)$$

where Qt is the outflow rate at time t, Q0 is the outflow rate at the start of the recession, and k is a rate constant. For this exponential decay model, the log transform of Qt (log Qt) is linearly related to time. Visual inspection of log Qt (Fig. 4.2b, segment 2-5) was performed to identify the point of inflection at which the behavior of log Qt remained linear up to the end of the storm response (Fig. 4.2b, segment 4-5). This segment of the hydrograph represents outflows that are composed of the more slowly released groundwater discharges after the more quickly discharged DR has exited the watershed as outflow. Linear regression results for this segment (Fig. 4.2b, segment 4-5) are extended back to the time of peak outflow (Fig. 4.2b, segment 3-4) to complete the separation of log Qt of the recession limb, and this is transformed back from log scale for the hydrograph separation. The peak BF (Fig. 4.2a, point 3) is connected to the point of initial rise in the hydrograph (Fig. 4.2a, point 1) to complete the separation into BF and DR. The quick response of the water table to rainfall creates an initial increase in BF as well as DR in the rising limb of the hydrograph. This is different from typical upland watersheds, where groundwater is less of an immediate influence on outflow and the hydraulic gradient is away from the stream. Total outflow volume and BF volume are calculated by taking the integrals under the hydrograph and hydrograph separation curve respectively. DR depth is calculated as

$$\frac{(Total\ outflow\ volume - Base\ flow\ volume)}{Watershed\ area} \tag{4-2}$$

The ROC is then calculated as the ratio of DR depth to rainfall depth.

For determination of the TSR, a start and end point of the hydrograph were defined. The hour just prior to the initial rise of the hydrograph for the storm event was chosen as the start point (Fig. 4.2a, point 1). The hydrograph was extended until outflow rates returned to the initial outflow rate at the start point to signify the end of the storm response (Fig. 4.2a, point 5). Total storm-flow depth is represented in Fig. 4.2a as the sum of areas A and B and it is defined as

$$\frac{Total\ outflow\ volume - (Duration\ of\ storm\ response * Initial\ Outflow\ rate)}{Watershed\ area} \tag{4-3}$$

The TSR is then calculated as the ratio of total storm-flow depth to rainfall depth. For some storm events, the recession limb was interrupted by additional rainfall before outflows returned to the initial outflow level. In these cases, the BF recession that was calculated by hydrograph separation for the storm was used to model outflows outward until the antecedent value was met in order to model the total storm response to rainfall.

Storm event rainfall and measures of AMC were compared to ROC and TSR values by linear regression to assess the relationship between AMC and runoff generation. Storm events were separated by growing season dates for the respective watershed locations to determine seasonal trends in mean ROC and TSR. An F-test was used to determine if the seasonally grouped storm events had equal variance, and the appropriate two-sample t-test for equal or unequal variance was used to determine if there was a difference between mean values of ROC and TSR. Segmented regression was performed using SegReg, a program designed to calculate a break point in a dataset joining segments of separate linear regression that improves upon simple linear regression estimates for the data. This program was developed according to statistical principles outlined by Oosterbaan et al. (1990). Water table break points estimated by segmented regression were subsequently used to separate storm events between dry and wet AMC in order to assess differences in runoff generation, and mean ROC and TSR were analyzed in the same manner used to assess seasonal mean differences. The difference between the peak outflow rate and the initial outflow rate for storm events was compared between the two watersheds to assess the difference in magnitude of peak outflow rates in the rainfall response on the watersheds. The mean values were compared using F-tests and the two-sample t-test for mean comparison as well.

RESULTS AND DISCUSSION

On the UDC watershed, 23 storm events met the selection criteria for analysis. Twenty storm events were analyzed on WS80. Rainfall depth, initial outflow rate, 5- and 30- day API, initial water table elevation, peak outflow rate, total storm-flow depth, direct runoff depth as estimated by hydrograph separation, TSR, and ROC were compiled for all storms to assess outflow and direct runoff generation as it relates to AMC. These data are summarized in Table 4.1 for UDC storm events and Table 4.2 for WS80 storm events.

Table 4.1. Summary of Upper Debidue Creek storm event characteristics for a three-year period.

Date	P	Ant. Q	API5	API 30	WT	Peak Q	DR	SF	ROC	TSR
7/24/2008	30	17	11	250	NA	112	1	5	0.02	0.18
9/5/2008	87	42	0	150	NA	1,102	12	47	0.14	0.54
9/11/2008	25	166	81	231	NA	640	4	23	0.18	0.93
9/16/2008	47	159	0	171	NA	904	12	30	0.25	0.64
9/25/2008	42	134	1	175	NA	704	10	33	0.23	0.78
3/1/2009	40	16	1	33	3.19	428	3	34	0.08	0.83
4/2/2009	60	83	18	62	3.38	890	9	56	0.16	0.93
8/28/2009	68	0	0	35	2.34	99	0	3	0.00	0.04
11/10/2009	78	0	5	74	2.42	167	2	14	0.03	0.18
1/16/2010	22	119	0	74	3.60	408	2	20	0.11	0.89
1/25/2010	23	253	17	60	3.73	673	5	15	0.24	0.68
2/2/2010	27	311	20	82	3.74	896	7	22	0.25	0.82
3/2/2010	24	158	0	129	3.61	462	8	21	0.32	0.88
5/4/2010	23	2	0	30	3.25	40	0	1	0.00	0.05
6/20/2010	36	0	0	81	2.85	27	0	0	0.00	0.00
6/30/2010	35	0	36	140	2.93	64	0	1	0.00	0.02
7/10/2010	35	0	19	139	2.95	75	0	2	0.01	0.04
8/1/2010	24	0	6	130	3.03	31	0	0	0.01	0.02
8/13/2010	40	23	40	149	2.97	376	1	2	0.03	0.04
8/19/2010	25	2	36	143	3.29	85	0	2	0.01	0.09
6/29/2011	20	0	0	37	2.49	47	0	1	0.01	0.05
8/6/2011	81	24	6	179	2.66	656	3	18	0.04	0.22
8/25/2011	67	51	13	117	2.76	234	5	10	0.08	0.15

P = rainfall (mm), Ant. Q = initial outflow (m^3/hr), API = antecedent precipitation index (5- and 30-day, mm), WT = initial water table elevation (m asl), Peak Q = peak outflow (m^3/hr), DR = direct runoff (mm), SF = total storm flow (mm)

Table 4.2. Summary of watershed (WS80) storm event characteristics for a 3-year period.

Date	P	Ant. Q	API 5	API 30	WT	Peak Q	DR	SF	ROC	TSR
8/21/2008	37	14	31	171	8.94	197	1	4	0.02	0.11
9/5/2008	98	0	1	167	8.26	862	2	13	0.02	0.13
9/9/2008	113	31	98	255	8.97	5,298	31	54	0.28	0.48
9/25/2008	65	0	1	218	8.62	634	5	15	0.08	0.23
10/24/2008	154	2	0	156	8.79	13,424	88	115	0.57	0.74
11/29/2008	47	2	0	27	8.64	217	4	10	0.09	0.22
3/1/2009	58	6	4	36	8.80	1,446	14	25	0.23	0.43
4/2/2009	67	49	35	61	9.02	2,619	23	46	0.34	0.69
7/16/2009	41	1	30	122	8.69	117	1	2	0.01	0.05
7/22/2009	29	1	0	153	8.80	54	0	1	0.01	0.03
8/31/2009	57	0	48	94	7.85	303	1	3	0.02	0.06
11/11/2009	70	0	2	61	7.53	22	0	1	0.00	0.01
12/18/2009	67	40	12	136	9.06	2,691	26	42	0.39	0.62
12/25/2009	31	77	3	200	9.06	967	7	18	0.22	0.57
1/16/2010	51	13	0	114	8.97	1,713	9	36	0.17	0.70
1/25/2010	42	167	25	88	9.08	2,119	19	29	0.46	0.69
3/28/2010	31	14	0	81	8.97	387	4	11	0.14	0.34
5/4/2010	52	0	0	21	8.07	13	0	0	0.00	0.01
9/29/2010	75	102	147	188	9.09	2,944	12	30	0.15	0.40
2/2/2011	66	4	12	60	8.88	231	10	14	0.15	0.22

P = rainfall (mm), Ant. Q = initial outflow (m^3/hr), API = antecedent precipitation index (5- and 30-day, mm), WT = initial water table elevation (m asl), Peak Q = peak outflow (m^3/hr), DR = direct runoff (mm), SF = total storm flow (mm)

Histograms of the storm event rainfall sizes on both watersheds show storm size distributions (Fig. 4.3) over the three-year period. The annual pattern of rain event sizes on both watersheds is presented in Figure 4.4 for storm events plotted by Julian date over the three years of data. Storm event sizes on both watersheds ranged from 20 – 154 mm with larger storms during the fall season months coinciding with tropical storms.

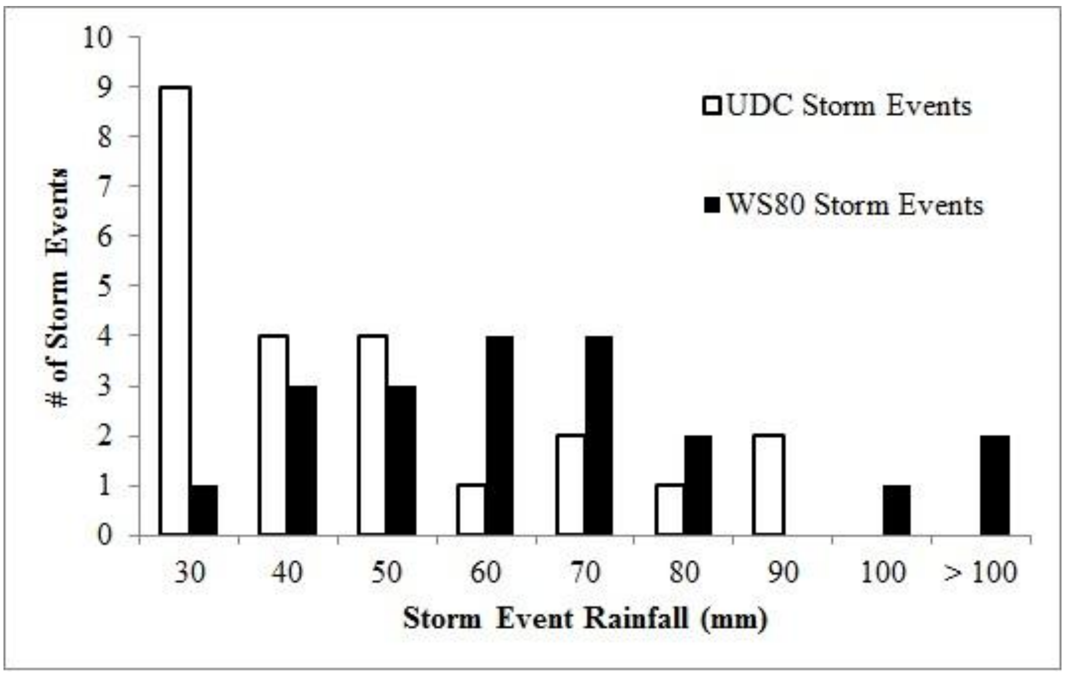

Figure 4.3. Histograms of storm event sizes for Upper Debidue Creek and Watershed 80 watersheds.

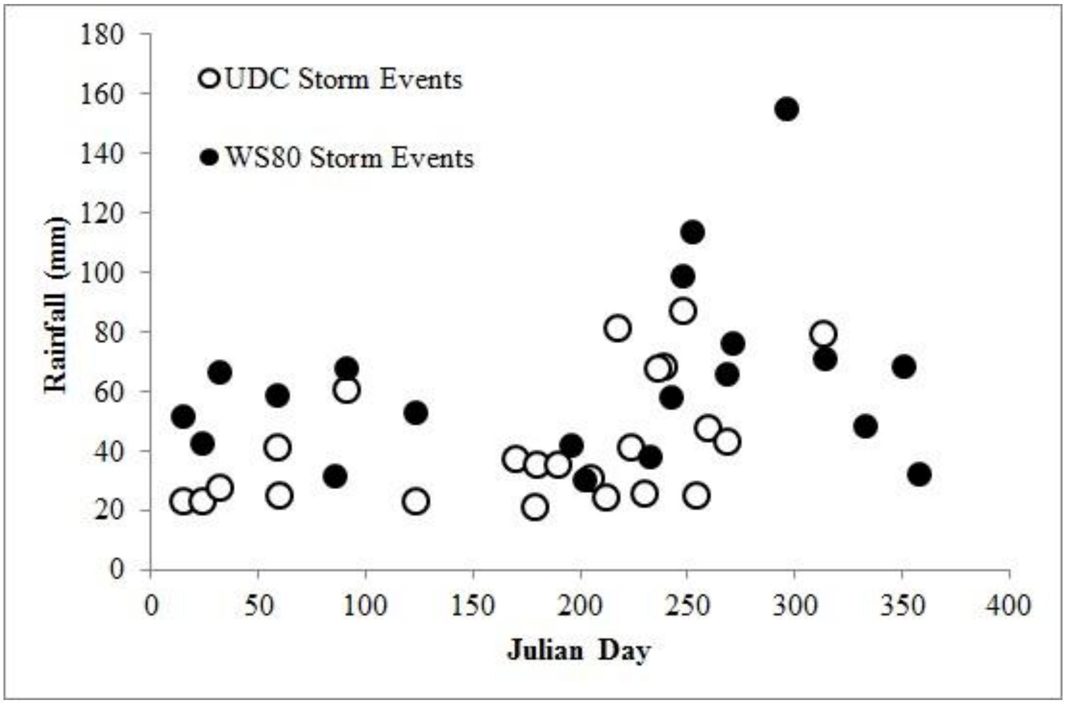

Figure 4.4. Storm event distribution by Julian day for UDC and WS80 storms.

The TSR and ROC coefficients for storms on the two watersheds varied widely as expected. The ROC showed a range of 0 to 0.32 with a mean of 0.10 at UDC and from 0 to 0.57 with a mean of 0.17 at WS80 (Table 4.3). The TSR varied from 0 to 0.93 at UDC with a mean of 0.39 and from 0.01 to 0.74 with a mean of 0.34 at WS80. Although the two means of the TSR were similar the calculated COV of 0.76 on WS80 shows a less variability of storm events than on UDC watershed with a COV of 0.95.

Table 4.3. Descriptive statistics for UDC and WS80 storm events overall and separately for dormant season and growing season events.

UDC	**Overall**		**Dormant Season**		**Growing Season**			
	ROC	TSR	ROC	TSR	ROC	TSR		
Mean	0.10	0.39	0.20	0.82	0.07	0.27		
Median	0.04	0.18	0.24	0.83	0.03	0.12		
Std. Dev.	0.10	0.37	0.10	0.08	0.08	0.33		
Min	0.00	0.00	0.08	0.68	0.00	0.00		
Max	0.32	0.93	0.32	0.89	0.25	0.93		
Count (n)	23	23	5	5	18	18		
COV	1.06	0.95	0.51	0.10	1.25	1.21		
WS80	**Overall**		**Dormant Season**		**Growing Season**		**Growing Season w/o 20081024**	
	ROC	TSR	ROC	TSR	ROC	TSR	ROC	TSR
Mean	0.17	0.34	0.24	0.49	0.13	0.25	0.09	0.21
Median	0.15	0.28	0.22	0.57	0.02	0.13	0.02	0.12
Std. Dev.	0.17	0.26	0.13	0.21	0.17	0.26	0.12	0.22
Min	0.00	0.01	0.09	0.22	0.00	0.01	0.00	0.01
Max	0.57	0.74	0.46	0.70	0.57	0.74	0.34	0.69
Count (n)	20	20	7	7	13	13	12	12
COV	0.99	0.78	0.54	0.42	1.36	1.02	1.28	1.05

The range in ROC and TSR values were hypothesized to coincide with seasonal trends in AMC. Intermittent outflow and low runoff generation is typical during the summer months as shown in Tables 4.1 and 4.2 for June, July, and August storm events. Higher groundwater elevations contribute to sustained BF conditions and higher runoff generation during the winter months of December, January and February (Tables 4.1 and 4.2). This trend can be visualized in the measured ROC values plotted by Julian day over the three-year period at UDC and WS80 watersheds (Fig. 4.5a and 4.5b).

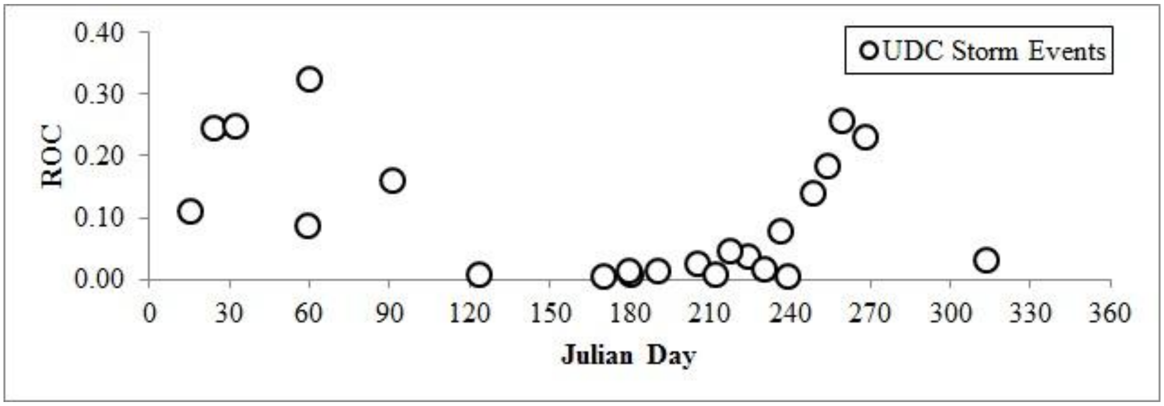

Figure 4.5a. Runoff coefficients plotted by Julian day for Upper Debidue Creek storm events for the three-year period.

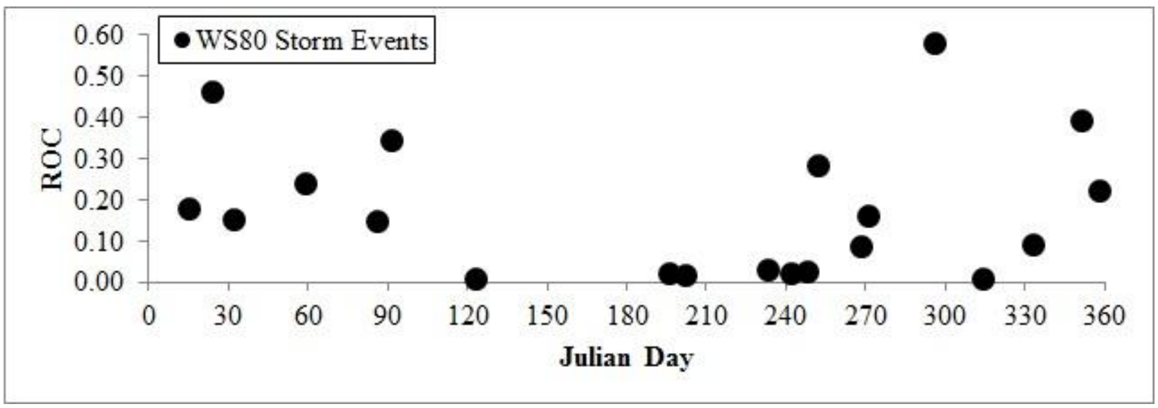

Figure 4.5b. Runoff coefficients plotted by Julian day for Watershed 80 storm events for the three-year period.

Mean ROC and TSR were higher on both watersheds during the dormant season. Table 4.3 summarizes select descriptive statistics for each watershed across all storm events and separately for the dormant season and growing season events. At UDC the mean TSR was 0.82 for the dormant season. This was significantly higher ($p < 0.01$) than 0.27 for the growing season with high evaporative demands. At WS80 also, the mean TSR of 0.49 in the dormant season was significantly higher ($p = 0.024$) than 0.25 for the growing season. This reflects the role of seasonal trends in evapotranspiration and soil moisture conditions on runoff generation in coastal headwater streams as was shown by La Torre Torres et al. (2011). Similarly, the mean ROC had similar seasonal differences on the two watersheds. At UDC, the dormant season mean ROC of 0.20 was higher than the 0.07 mean ROC measured for growing season storm events ($p < 0.01$). There was no difference in mean ROC between dormant (0.24) and growing season (0.13) storm events on WS80 at the 95% confidence level ($p = 0.07$). The tropical storm event that occurred on October 24, 2008 during the growing season at WS80 measured 154 mm of rain with an ROC of 0.57. This was the largest storm analyzed during the study period. The mean ROC at WS80 for the growing season events was 0.13 and the median value was 0.02. This difference is due to the effect of the very high ROC value for this storm. By omitting the storm event for October 24, 2008 at WS80, the dormant season mean ROC of 0.24 was higher than the revised growing season mean ROC of 0.09 ($p < 0.01$). Large tropical storms that occur at the end of the growing season have the potential to produce large amounts of runoff that deviate from more typical growing season trends, mainly due to high amount of rainfall and tapered ET demands during late fall months of October and November. Runoff generation for these large, high intensity tropical storms is more closely related to rainfall characteristics than seasonal trends of AMC. Previous studies have shown that storm-event outflows and runoff generation on these LCP headwater catchments are not well predicted by rainfall alone (Harder et al., 2007; La Torre Torres et al., 2011). The relationships between rainfall and runoff generation measured by linear regression are summarized in Table 4.4.

Table 4.4. Coefficient of determination for the relationship between rainfall and ROC and TSR for UDC and WS80 storm events.

		UDC	**WS80**			**UDC**	**WS80**
ROC	R^2	0.01	0.23	TSR	R^2	0.01	0.10
	p-value	0.64	0.03		p-value	0.68	0.18

This relationship was only significant for the ROC at WS80 ($r^2 = 0.23$, $p = 0.03$). There was no significant ($\alpha = 0.05$) relationship between rainfall and ROC at UDC or between rainfall and TSR at either watershed. Runoff generation does not have a strong relationship to rainfall due to the variable AMC on LCP headwater streams. Variable AMC produces a range of ROC and TSR for similar rainfall depths due to differences in runoff generation mechanisms.

Antecedent Moisture Conditions

The results of linear regressions performed to assess the relationships between measures of AMC and runoff generation are summarized in Table 4.5.

Table 4.5. Summary of linear regression results for relationships between AMC parameters and ROC and TSR and between water table and initial outflow.

ROC Regressions		**Initial Outflow**	**API-5**	**API-30**	**Water Table**
UDC	R^2	0.78	0.00	0.02	0.54
	p-value	$p < 0.01$	0.93	0.52	$p < 0.01$
WS80	R^2	0.25	0.01	0.01	0.31
	p-value	0.03	0.78	0.66	0.01
WS80*	R^2	0.50	0.03	0.00	0.42
	p-value	$p < 0.01$	0.46	0.86	0.00
TSR Regressions		**Initial Outflow**	**API-5**	**API-30**	**Water Table**
UDC	R^2	0.56	0.01	0.00	0.53
	p-value	$p < 0.01$	0.71	0.86	$p < 0.01$
WS80	R^2	0.29	0.01	0.03	0.45
	p-value	0.01	0.69	0.50	0.00
WS80*	R^2	0.41	0.02	0.02	0.49
	p-value	0.00	0.52	0.62	$p < 0.01$
WS80* = Analysis performed omitting the storm on 20081024					

Analysis was performed for WS80 storm events twice to determine the effect that the event on October 24, 2008 had on calculations. This large tropical storm event measured 154 mm of rain that mostly fell over a 13 hour period with an intensity of nearly 12 mm/hr. Runoff generation for this storm is more closely related to rainfall characteristics than the AMC, so it has a biasing effect on the data. Results that include this event in the analysis will be presented for WS80 and those that omit it will be noted accordingly. There was no significant relationship between API and ROC or TSR using the 5- or 30-day API measure. It does not appear that antecedent precipitation is a good measure to estimate the AMC as it relates to runoff generation for these watersheds as was also recently noted by La Torre Torres et al. (2011). The relationship between initial outflow and the ROC and TSR was significant at both locations, consistent with Amatya et al. (2000) who used the initial outflow rate as the AMC in their study. At UDC, initial outflow rate explained a larger percentage of the variability for ROC ($r^2 = 0.78$, $p < 0.01$) and TSR ($r^2 = 0.56$, $p < 0.01$) than at WS80 ($r^2 = 0.25$, $p = 0.03$ for ROC; $r^2 = 0.29$, $p = 0.01$ for TSR). The relationship at WS80 was greater when the October 24, 2008 storm was omitted for both ROC ($r^2 = 0.50$, $p < 0.01$) and TSR ($r^2 = 0.41$, $p < 0.01$). Outflow in these headwater streams is a result of gravity-driven groundwater drainage and storm-event surface runoff. The magnitude of initial outflow shows a significant ($\alpha = 0.05$) relationship to runoff generation as a measure of the AMC for these streams. Because outflows are influenced by groundwater elevation, the initial water table elevation also shows a significant relationship to runoff generation. A larger portion of the variability in runoff coefficients was explained by the water table elevation as AMC at UDC (r^2= 0.54, $p < 0.01$ for ROC; $r^2 = 0.53$, $p < 0.01$ for TSR) than at WS80 ($r^2 = 0.31$, $p = 0.01$ for ROC; $r^2 = 0.45$, $p < 0.01$ for TSR). This relationship improved at WS80 again with the omission of the October 24, 2008 storm for ROC ($r^2 = 0.42$, $p < 0.01$) and TSR ($r^2 = 0.49$, $p < 0.01$). Because sustained BF that is not directly influenced by rainfall on these headwater streams consists of groundwater discharges, initial outflows were compared to initial water table elevation to determine how closely they were related. A significant relationship was measured for both the watersheds. The initial water table elevation accounted for more of the variability in initial outflow measured on the UDC watershed ($r^2 = 0.56$, $p < 0.01$) than at WS80 ($r^2 = 0.24$, $p = 0.03$). This may be partially due to differences in well placement between the watersheds and differences in topography. The much lower gradient at UDC is likely to influence a closer relationship between groundwater fluctuations and streamflow dynamics. These results demonstrate a significant relationship between water table elevation and outflow rates that is consistent with the relationship between water table elevation and runoff generation in these headwater streams. The 5- and 30-day API did not appear to have a direct relationship with runoff generation, but they likely influence water table elevations indirectly by infiltrative replenishment of the groundwater aquifer subject to evapotranspirative demands. There was a stronger relationship between initial outflow and water table elevation with the ROC and TSR coefficients at UDC than at WS80. Outflows and runoff generation may be more closely related to varying water table elevations at this watershed as also affected by microtopography and surface storage.

Influence of Water Table Elevation

Initial outflow rate, ROC, and TSR all display a similar non-linear relationship to initial water table elevation on both the watersheds (Fig. 4.6 and 4.7).

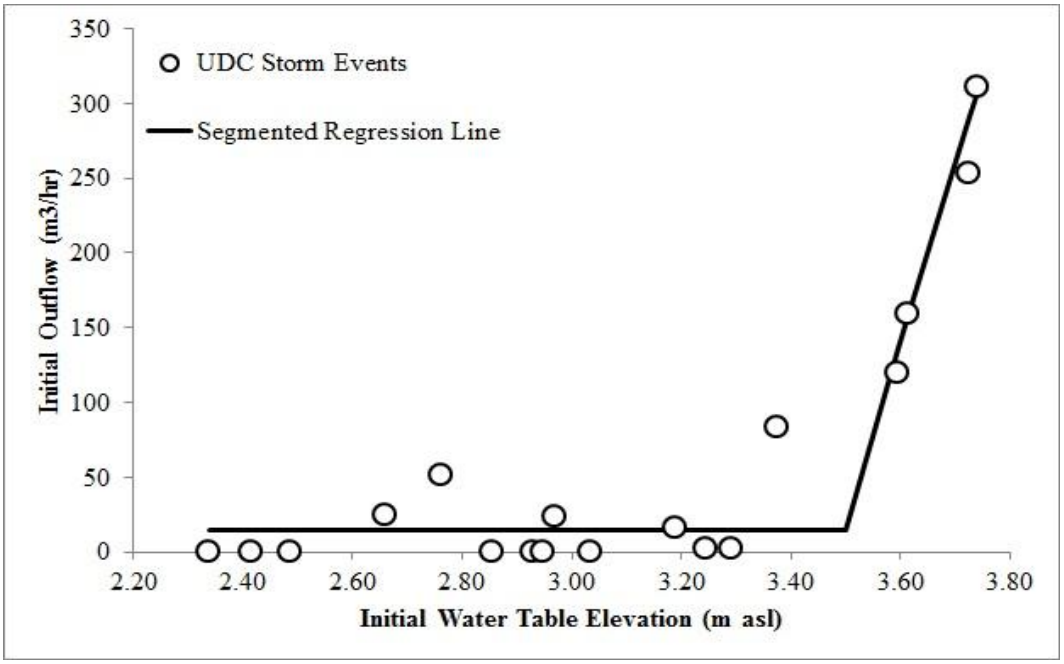

Figure 4.6a. Initial outflow rate plotted against initial water table elevation with segmented regression trend line for Upper Debidue Creek.

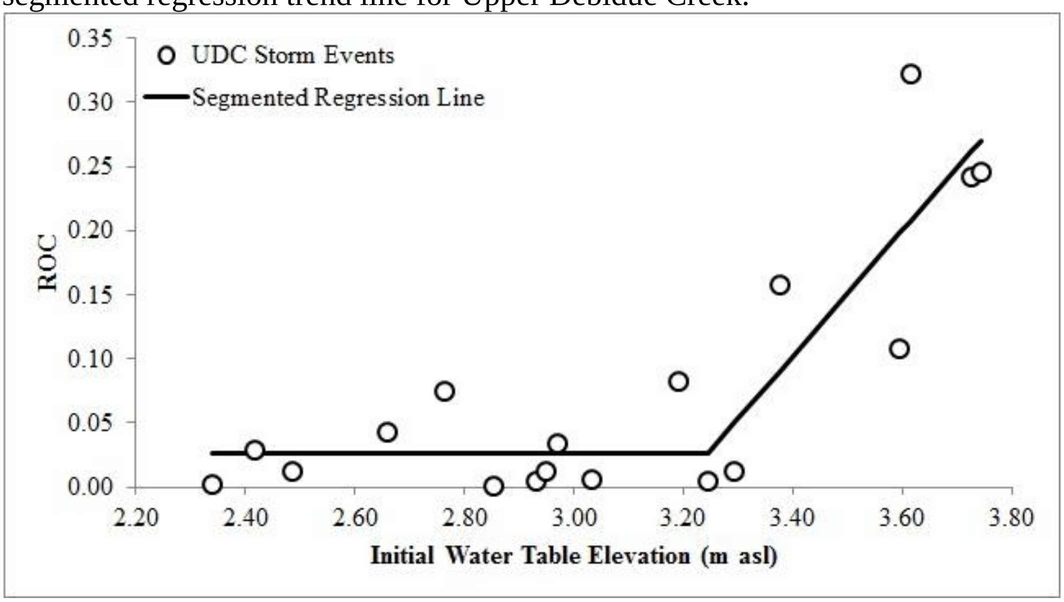

Figure 4.6b. Runoff coefficients plotted against initial water table elevation with segmented regression trend line for Upper Debidue Creek.

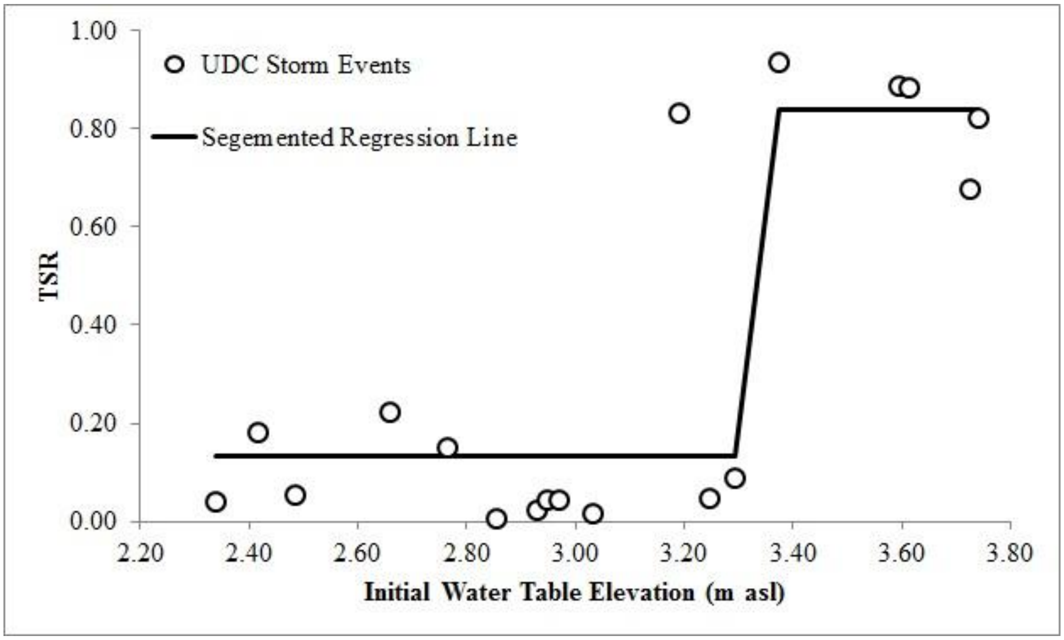

Figure 4.6c. Total storm response plotted against initial water table elevation with segmented regression trend line for Upper Debidue Creek.

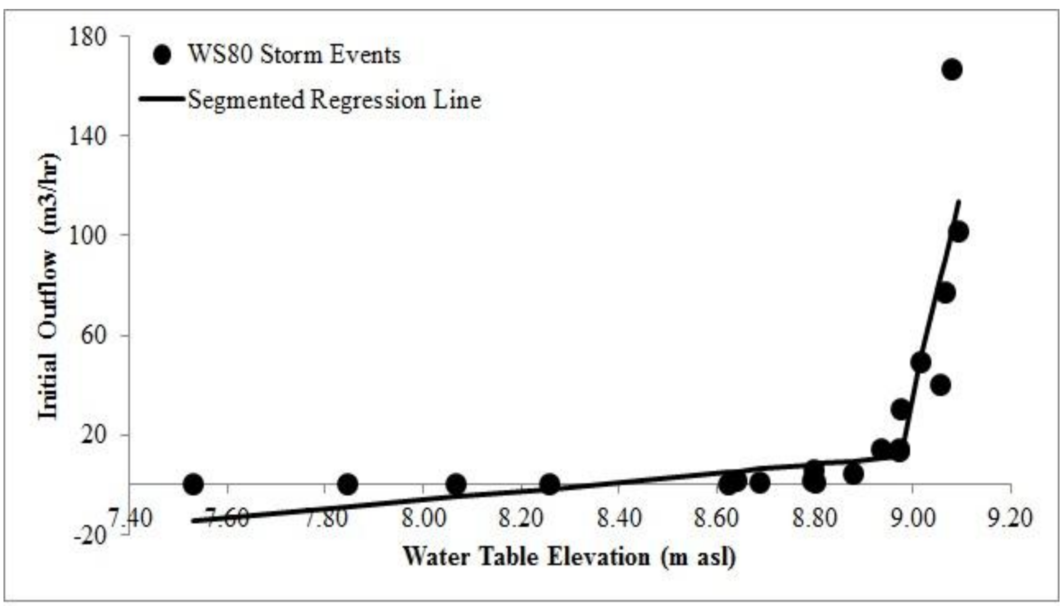

Figure 4.7a. Initial outflow rate plotted against initial water table elevation with segmented regression trend line for Watershed 80.

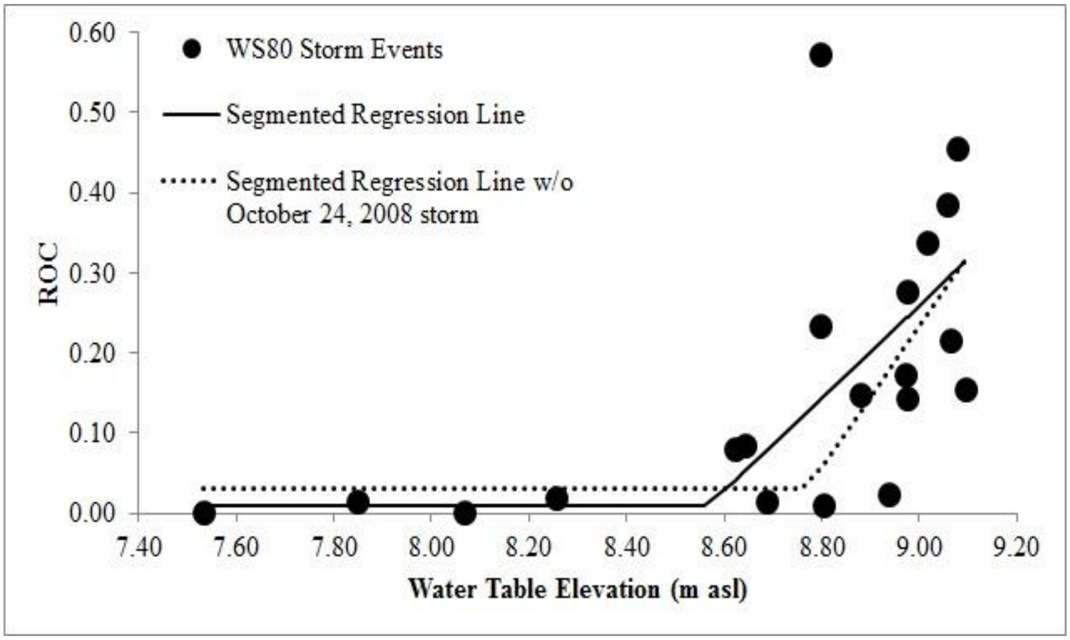

Figure 4.7b. Runoff coefficients plotted against initial water table elevation with segmented regression trend line for Watershed 80.

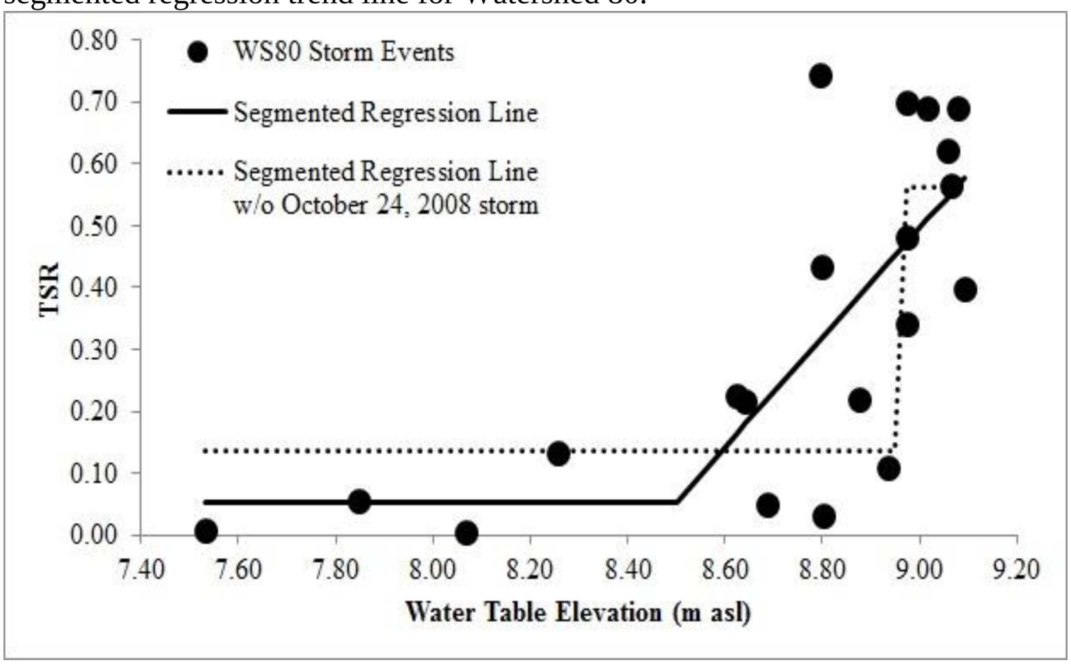

Figure 4.7c. Total storm response plotted against initial water table elevation with segmented regression line for Watershed 80.

Lower watershed outflows are associated with low water table elevations prior to rainfall and low surface runoff generation that remains fairly constant as water table elevation increases to the surface after complete saturation. There is a threshold groundwater elevation for each watershed that is approximately consistent for initial outflow rate, ROC, and TSR at which the relationship between the water table elevation and runoff generation diverges (Figs.4. 6 and 4.7). Groundwater elevations above this level display an approximate linear increase in outflow and runoff generation. Segmented regression results are summarized in Table 4.6 and show an average break point of 3.35 m at UDC and 8.68 m at WS80.

Table 4.6. Summary of segmented regression results for initial outflow rate, ROC, and TSR against water table elevation.

		Initial Outflow	ROC	TSR	Avg. Break Point (m asl)
UDC	Break Point (m asl)	3.5	3.25	3.29	3.35
	R^2	0.936	0.778	0.745	
	p-value	$p < 0.01$	$p < 0.01$	$p < 0.01$	
WS80	Break Point (m asl)	8.97	8.56	8.5	8.68
	R^2	0.8	0.342	0.528	
	p-value	$p < 0.01$	$p < 0.01$	$p < 0.01$	
WS80*	Break Point (m asl)	NA	8.76	8.95	8.89
	R^2	NA	0.582	0.734	
	p-value	NA	$p < 0.01$	$p < 0.01$	

This groundwater elevation is 0.84 m below ground surface (bgs) at UDC and 0.4 m bgs at WS80 at the respective measuring wells. These results support the findings of

Williams (2007) who observed that groundwater elevations below a similar threshold level (approximately 1 m bgs at the watershed divide for the coastal watershed studied near Georgetown, SC) were associated with high soil storage capacity and low runoff generation. For groundwater elevations above this level, both the outflow and runoff generation increased with water table elevation. Using 2003-04 data Harder et al. (2007) also found a significant ($\alpha = 0.05$) non-linear power relationship between water table and outflow generation on the WS80 watershed. The relationship between water table elevation and initial outflow rate, ROC, and TSR are all improved when modeled by a segmented regression for both the watersheds. These results (summarized in Table 4.6 and displayed graphically in Figs. 4.6 and 4.7) demonstrate the significance of this break point water table elevation in predicting outflow and runoff generation on these LCP headwater streams. Results suggest that this groundwater elevation may have physical significance on these watersheds that is related to runoff generation. A break point water table elevation could dictate the rainfall response on these watersheds between dry and wet AMC. Rain events that occur when the water table elevation is below the break point could be expected to produce very moderate amounts of runoff for dry AMC. When the water table elevation is above the break point, substantial runoff generation would be expected given higher water table conditions. Storm events with water table elevation below the calculated break point were designated as dry AMC and those with water table elevation above the break point were designated as wet AMC. Descriptive statistics are summarized in Table 4.7.

Table 4.7. Summary of descriptive statistics for storm events separated between dry and wet AMC according to water table elevation.

UDC	**Dry Antecedent**		**Wet Antecedent**	
	ROC	*TSR*	*ROC*	*TSR*
Mean	0.02	0.13	0.22	0.84
Standard Error	0.01	0.06	0.04	0.04
Median	0.01	0.05	0.24	0.88
Standard Deviation	0.03	0.22	0.08	0.10
Minimum	0.00	0.00	0.11	0.68
Maximum	0.08	0.83	0.32	0.93
Count (n)	13	13	5	5
WS80	**Dry Antecedent**		**Wet Antecedent**	
	ROC	*TSR*	*ROC*	*TSR*
Mean	0.03	0.11	0.23	0.43
Standard Error	0.02	0.04	0.04	0.07
Median	0.02	0.10	0.19	0.46
Standard Deviation	0.04	0.10	0.17	0.25
Minimum	0.00	0.01	0.01	0.03
Maximum	0.09	0.23	0.57	0.74
Count (n)	6	6	14	14

The mean TSR for dry AMC storms was significantly lower ($\alpha = 0.05$) than that for wet AMC storms at UDC ($p < 0.01$) and at WS80 ($p < 0.01$). Mean ROC was similarly lower for dry AMC storms than wet AMC storms at UDC ($p < 0.01$) and at WS80 ($p < 0.01$). This is similar to the observed differences in seasonal runoff generation and highlights seasonal trends in AMC. Storms that are categorized as wet or dry AMC according to the water table elevation were observed during both the dormant and growing season, but the majority of dry AMC storm events are seen during the growing season and wet AMC storms during the dormant season. Variability in rainfall patterns from year to year and also on a seasonal basis can produce variable moisture conditions on these watersheds during months between periods of high and low evapotranspiration as the groundwater aquifer (as it relates to soil water storage) is depleted during the spring months or replenished in the fall. This is observed in variable runoff generation at UDC and WS80 during these transitional months (Fig. 4.5a and 4.5b) between the wetter winter months and the drier summer months.

It is hypothesized that the break point water table elevation at each watershed relates to runoff generation mechanisms that are unique to low topography LCP hydrology and are defined by low gradient, groundwater influence, and soil characteristics. Differences between clayey soils and sandy soils in soil storage capacity, infiltration, and hydraulic conductivity influence runoff generation dynamics (La Torre Torres et al., 2011) and contribute to differences between breakpoint water table elevations at the two locations because of the effect that these soil properties have on the movement of runoff and groundwater towards the stream. The water table elevation must be sufficiently high for outflow to occur. When the water table elevation falls below a certain level, outflows cease on these intermittent streams due to disconnection from the groundwater aquifer. This is common during the summer months when evapotranspirative demand exceeds groundwater replenishment by rainfall infiltration, consistent with the observations noted by Todd et al (2006) who reported that some parts of the basin may become decoupled from the basin outlet as summer progresses and their runoff may be lost to evaporation and infiltration or held in surface storage even before reaching the outlet. Water table elevations below this level represent a dry AMC associated with higher soil storage capacity that has to be filled by infiltration form a large rainfall amount before runoff responses to the rainfall occur. As water table elevations increase above this level, higher outflow rates prior to rainfall are observed and runoff generation increases for storms that occur during periods of wet AMC. Saturation excess overland flow is the dominant source for runoff generation on low gradient forested watersheds in the LCP (Eshleman et al., 1994; Williams, 2007; Slattery et al., 2006). It has been postulated that as the groundwater elevation increases, the area of saturation near the stream increases and larger areas contribute to saturated excess overland flow, producing greater runoff (Eshleman et al., 1994). Accelerated groundwater contributions to outflow are also believed to increase as the water table elevation rises above this break point level due to a greater hydraulic gradient toward the stream and the potential for piston flow discharges (Williams 2007). The differences between BF apportionment between the TSR and ROC coefficients highlight differences in runoff generation on these two watersheds.

Accounting for Base Flow

TSR and ROC are both measures of the rainfall response that differ by the way that they account for BF contributions as a portion of outflow. TSR measurements model BF as a constant and assign storm event runoff as all outflows in excess of the initial level. ROC measurements model BF as accelerated storm-event groundwater contributions that increase with peak outflows due to rapid water table rise and discharge according to a sustained recession that approximates natural groundwater discharges. It is difficult to directly measure groundwater contributions to storm-driven outflow, and the distinction between DR and BF is not easily made. Linear regression results between TSR and ROC demonstrate that the two measures of runoff share a good deal of agreement for storm events at UDC ($r^2 = 0.71$, $p < 0.01$) and at WS80 ($r^2 = 0.81$, $p < 0.01$). The relationship between TSR and ROC at each watershed approximates the additional DR that is calculated by the ROC model as the watershed produces more total outflow as modeled by the TSR method of runoff measurement (Fig. 4.8).

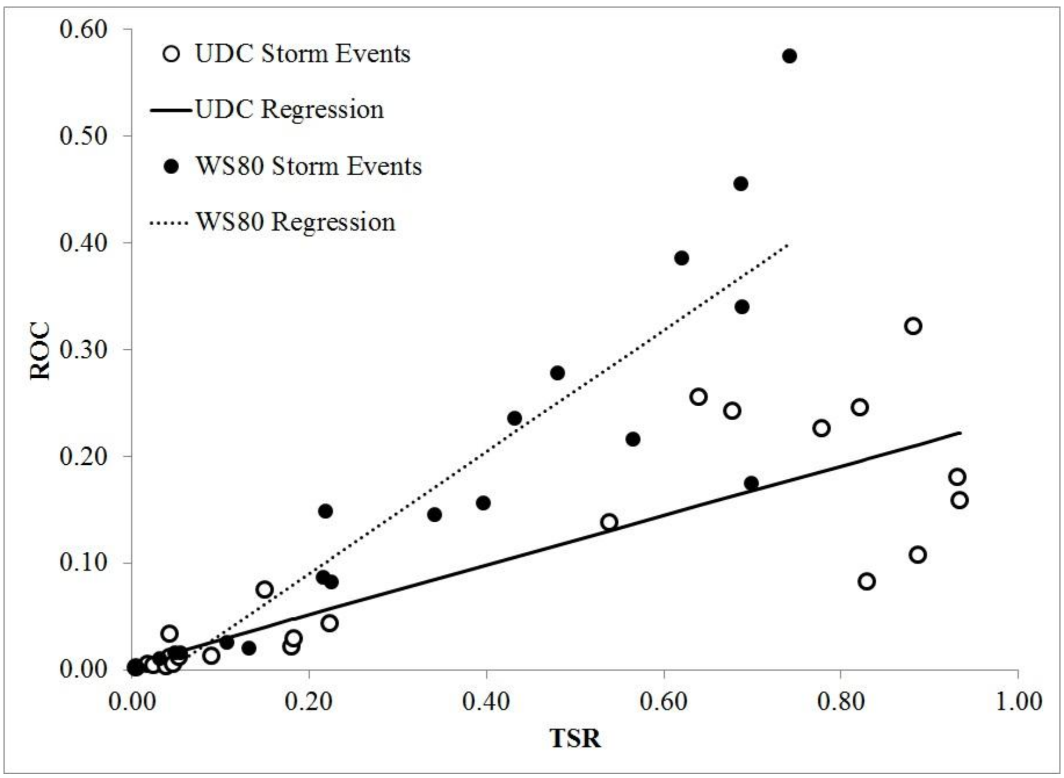

Figure 4.8. Runoff coefficient plotted against total storm response for Upper Debidue Creek and Watershed 80 storm events.

The slope of the trend line at UDC was 0.23 (p-value < 0.01), less than half that found for WS80, where the slope of the trend line was 0.57 (p-value < 0.01). Estimates of direct runoff at WS80 grow more than twice as fast than at UDC with increasing TSR. Additionally, the difference between peak outflow rate and the initial outflow rate were higher on average at WS80 than at UDC ($p = 0.01$). Higher peak outflows and greater direct runoff generation at WS80 are likely caused by clayey soils that are less infiltrative than sandy soils at UDC and generate greater surface-driven runoff in response to rainfall.

Mean ROC and TSR were similar at both watersheds for dry AMC storm events (Table 4.7). Mean ROC was 0.02 at UDC and 0.03 at WS80, and mean TSR was 0.13 at UDC and 0.11 at WS80. No significant difference in mean ROC or mean TSR was found between UDC and WS80 ($\alpha = 0.05$). BF levels for dry AMC storms are low due to smaller groundwater contribution. The similar results for mean ROC and TSR at UDC and WS80 reflect similar runoff generation mechanisms on both watersheds for dry conditions. Mean ROC was similar on both watersheds for wet AMC storm events, measuring 0.22 at UDC and 0.23 at WS80. No significant difference between the means was measured ($\alpha = 0.05$). While the ROC model measures similar DR contributions to outflow for both the watersheds under wet conditions, the same is not true for TSR. Mean TSR for wet AMC storms at UDC (0.84) was significantly greater than at WS80 (0.43) by comparison ($p < 0.01$). Under wet conditions with high groundwater elevation, model results demonstrate much larger total outflow generation in the rainfall response for the UDC watershed but ROC estimates are similar for both watersheds. The TSR model for runoff estimates does not account for accelerated groundwater contributions in the rainfall response. The difference in results between the two watersheds based on mean TSR indicates that this BF component is greater at UDC than at the WS80. Groundwater contributions to outflow may not be directly attributable to new rainfall inputs for a storm event, especially when the groundwater elevation is higher during wet AMC storm events. There is a stronger relationship between water table elevation and runoff generation measures at UDC than at WS80 (Table 4.5). This relationship indicates that groundwater-driven BF contributions and runoff generation are more closely tied at UDC. Runoff generation for similar watersheds is better estimated by the ROC model because it accounts for accelerated BF contributions. The relationship between water table elevation and runoff generation is also strong at WS80 as was also shown earlier by Harder et al. (2007). Higher direct runoff generation and higher peak outflow rates may be the result of clayey soils that generate additional surface-driven runoff. Clayey soils are typically assigned to Hydrologic Soil Group D (NRCS, 1986) which represents the group of soils associated with the highest runoff generation for the purpose of Curve Number method runoff modeling. These site-specific soil characteristics may explain the relationship between rainfall and ROC that was observed (Table 4.4). Groundwater elevations display a significant ($\alpha = 0.05$) relationship with ROC and TSR for both of these LCP headwater streams, and seasonal trends in antecedent moisture levels related to the groundwater elevation determine the rainfall response.

CONCLUSIONS

The runoff response to rainfall on LCP headwater streams is not easy to characterize due to variable moisture conditions and the range of other parameters affecting the soil moisture. Runoff generation cannot be not predicted by rainfall alone for any given storm event, nor is antecedent precipitation a good indicator of the AMC. The influence of the groundwater aquifer on outflow generation seems to best determine the rainfall response. Outflow rate and the water table elevation at the start of rainfall characterize the AMC and are the best predictors of runoff generation. The relationship between the water table elevation, runoff amount, and event outflows on these watersheds appears to change at a given break point elevation. Water table elevations below this level define dry AMC, and low runoff generation in this condition typically results due to high soil storage and possibly disconnected surface (due to microtopography) as well as groundwater. Water table elevations above this level define wet AMC, and high runoff generation occurs due to saturated conditions. Stronger relationships between water table elevation and outflows at UDC suggest that groundwater and surface water generation may be more closely related on LCP headwater streams with sandy soils than on watersheds with clayey soils like WS80. For these watersheds with sandy soil, groundwater contributions to outflow represent a greater portion of streamflows, and runoff estimates may be better predicted by the ROC. Watersheds in the LCP with clayey soils also demonstrate a strong relationship between water table elevation and runoff generation, but surface-derived runoff may occur at a more substantial level on these watersheds due to less infiltrative soils. These results may have large implications on management of stormwater runoff and best management practices in the LCP landscapes.

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CHAPTER FIVE

ASSESSMENT OF THE CURVE NUMBER METHOD ON TWO HEADWATER STREAMS IN LOWER COASTAL PLAIN SOUTH CAROLINA

ABSTRACT

Increasing development in the Lower Coastal Plain (LCP) of South Carolina requires that better water budget calculations and predictions for runoff are made as land cover changes take place to ensure the continued ecological health and function of local watersheds. The low gradient topography and shallow water table that typify the area create unique hydrologic conditions that must be assessed and managed differently than higher gradient watersheds. The SCS Curve Number (CN) method is typically used for runoff estimation that determines the selection and design of stormwater management strategies and practices. The CN method was originally intended for small agricultural watersheds in higher gradient locations, and its applicability for LCP hydrological predictions has been questioned. Analyses of three years of storm event-based data for two LCP headwater catchments indicate that adaptations of method parameters must be made for reliable use. Back-calculated CNs from rainfall/runoff pairs ranged from 46 to 90 at the Upper Debidue Creek (UDC) watershed and from 42 to 89 at Watershed 80 (WS80). Curve-fitting results suggest that actual CNs for LCP headwater catchments are lower than values selected from available tables for pre-development conditions. Watersheds that contain dual-HSG soils (e.g. B/D) that are related to shallow water table conditions are not modeled well by a single CN. Variation in runoff generation between storm events is related to soil moisture conditions that are best defined by the water table

elevation. These wet and dry conditions are related to seasonal trends in evapotranspiration, and runoff generation is not well modeled under the average moisture condition that is normally associated with the use of the CN method. Event analysis also indicates that the initial abstraction for these watersheds is better approximated as 5% of the maximum potential retention.

BACKGROUND

Land development and associated changes in land cover alter site hydrology. The natural movement of rainfall by infiltration or as surface runoff towards the watershed outlet is impacted when surface characteristics are altered. Development in previously unaltered watersheds raises the degree of impervious cover and lowers infiltration by the addition of impermeable surfaces, removal or reduction of vegetative cover, and increased soil compaction. As a result, and if not properly managed, higher relative surface runoff is generated while groundwater replenishment by infiltration decreases. The rainfall response of developed watersheds is characterized by higher runoff volumes and higher peak runoff flows compared to the pre-development condition (Arnold and Gibbons, 1996). A larger portion of rainfall quickly exits the watershed, and flashy runoff conditions can have negative consequences downstream, such as erosion, flooding, and water quality impairment due to a reduced time of concentration. Sustained groundwater base flows (BF) also decrease due to reduced groundwater replenishment, and this can have an impact on ecological health downstream (Tang et al., 2005; Walsh et al., 2005). These conditions have been shown to increase with the degree of impervious cover in developed watersheds (Arnold and Gibbons, 1996; Blair et al., 2008). Stormwater regulations typically require on-site handling of surface generated runoff when development takes place. Site runoff is handled by application of best management practices that are designed to reduce post-development runoff with a target of pre-development levels. The most widely accepted method for calculation of pre- and post-development runoff in the rainfall response is the SCS Curve Number (CN) method (Ponce and Hawkins, 1996). The CN method was originally developed for use on small agricultural watersheds in the Midwest United States where conditions are very different from LCP watersheds. It has been used to model runoff on watersheds that have very different hydrologic conditions than those that the method was originally calibrated for (Ponce and Hawkins, 1996; Van Mullem et al., 2002). The purpose of this study is to assess the CN method for two headwater catchments in the Lower Coastal Plain (LCP) of South Carolina to evaluate storm event rainfall and runoff data in comparison to results obtained from CN method calculations.

The LCP of South Carolina has unique hydrologic conditions that differ from higher gradient watersheds. The area is characterized by very flat topography and a shallow water table. The water table position changes in response to seasonal trends in evapotranspiration (ET). Water table elevation has been shown to contribute to variable moisture levels that determine outflow production and thus the response to rainfall (Amatya et al., 2006; Harder et al., 2007; La Torre Torres et al., 2011; Sun et al., 2002; Williams, 2007). High ET during the summer months lowers the water table, increasing available soil storage. Runoff generation is lower during these months, and watershed outflows become intermittent on headwater streams when the water table is lowered sufficiently. Declining ET during the fall months results in replenishment of the groundwater aquifer that raises the water table position when rainfall occurs. During the winter months when ET is lowest, the water table position remains high, contributing to higher watershed outflows and higher runoff generation during these months. Sustained BF on these headwater streams is a result of groundwater contributions. Headwater streams in the LCP function as natural gravity-driven drainage for the groundwater aquifer (Amatya et al., 2006). Sustained outflows not related directly to the rainfall response are a function of the water table position. Therefore, these outflows vary on an annual basis accordingly.

Runoff generation in the rainfall response on headwater catchments in the LCP ranges according to these variable conditions as well. Annual outflow depths measured as a percentage of annual precipitation range from less than 10% to greater than 50% for LCP headwater streams (Amatya et al., 2006; Harder et al., 2007). Yet runoff generation is not defined by rainfall alone. Antecedent runoff conditions (ARC, sometimes referred to as antecedent moisture conditions but here referenced to reflect TR-55 terminology (USDA, 1986) that are determined by soil moisture characteristics prior to rainfall play a large role in determining runoff for any given storm event. Harder et al. (2007) observed very low runoff generation in response to storm events for dry ARC on a LCP headwater catchment during a year in which the annual outflow depth measured only 8% of precipitation that fell. Sun et al. (2002) observed a large increase in runoff generation between comparable storms falling on dry and wet ARC on a LCP headwater catchment in Florida. Storm event outflows as a percentage of rainfall averaged 8% for dry ARC and 58% for wet ARC. Higher runoff generation for wet ARC was attributed to decreased soil storage and more rapid movement of rainfall to the watershed outlet as a result. Seasonal trends in ET contribute to trends in ARC, with wet ARC persisting during winter months and dry ARC during summer months. Analysis of runoff generation for storm events on a LCP headwater stream displayed higher mean watershed outflows as a percentage of event precipitation for storms during the wet winter months (0.33) than the dry summer months (0.21) (La Torre Torres et al., 2011). This seasonal variability in outflow is a function of variable water table elevation and the role that the groundwater aquifer has in defining ARC in these LCP headwater catchments.

Runoff generation mechanisms in the LCP differ from upland catchments where higher gradients determine outflows (Sun et al., 2002). Saturation excess overland flow is the dominant mechanism in LCP forested headwater catchments due to high infiltration rates and low gradient conditions (La Torre Torres et al., 2011; Williams, 2007). This process is influenced by water table elevations that have been shown to respond rapidly to rainfall (Williams, 1978). Saturated areas that generate runoff are determined by groundwater elevation and these areas vary in size between storms according to ARC. Variable source areas contribute to differences in runoff generation between storm events in lowland watersheds (Hewlett and Hibbert, 1965; Eshleman et al., 1994). The relationship between groundwater elevation and surface water generation in response to rainfall may also contribute to accelerated BF contributions to outflow. Rapid water table rise at the watershed divide once rainfall begins is thought to increase subsurface groundwater discharges towards the stream (Williams, 2007). Groundwater discharges toward the stream can also be increased by the piston-flow mechanism when the capillary rise of groundwater extends above the ground surface elevation and rainfall transmits pressure through the aquifer towards the stream (Williams, 2007). Shallow water table hydrology presents challenges because it is difficult to directly measure the groundwater contributions to outflows.

Curve Number Methodology

The CN method (as described in USDA Technical Release 55 (TR-55), 1986) assigns CNs to land surfaces based on hydrologic soil group (HSG), cover type, treatment, hydrologic condition, and antecedent runoff condition (ARC). The CN is a function of maximum potential retention, and it can be interpreted as the degree of permeability for the corresponding land cover conditions. Curve numbers have a range from 0 – 100 representing conditions from infinite infiltration to fully impermeable, respectively. Typical observed values range from 40-98, however, though they may be lower for forested conditions (Van Mullem et al., 2002). Soil types are assigned to one of four hydrologic soil groups (A, B, C, or D) based on infiltration and hydraulic conductance properties, with A being the most permeable and D the most impervious. Wet soils are assigned dual HSGs (A/D, B/D, C/D). These soils are assigned as Group D in the undrained condition and are better modeled as the alternate HSG if adequately drained (USDA, 2007). Typical land cover types have been classified, and a range of CNs for each has been calculated across the spectrum of HSGs. The cover types are further broken down by treatment (method of land management) and hydrologic condition (Good, Fair, or Poor; based on runoff potential and typically measured by density of plant cover) where applicable. ARC is a measure of antecedent moisture and it accounts for the range in runoff response that can be expected from dry (CN-I) to wet (CN-III) conditions. Most CN applications use the average ARC (CN-II) for runoff estimates.

The CN is a transformation of the variable S (mm), which represents the potential maximum retention of rainfall by the land.

$$CN = \frac{25{,}400}{S + 254} \tag{5-1}$$

S can also be thought of as the greatest possible difference between rainfall (P, mm) and direct runoff (Q, mm) for any given storm event. Representative CNs for the combination of land cover conditions and soil composition for a site are weighted by respective area percentages to produce a composite CN. This CN is applied to the CN equation to predict the direct runoff to be expected from any given storm.

$$Q = \frac{(P - I_a)^2}{(P - I_a) + S} \quad where\ P \geq I_a,\ otherwise\ Q = 0 \tag{5-2}$$

The remaining variable in the CN method is Ia (mm), or the initial abstraction. This variable represents the portion of rainfall that does not produce direct runoff. Initial abstraction is a composite of canopy interception, infiltration, surface storage, and other losses deducted from rainfall before direct runoff is produced (USDA, 1986). The quantity ($P - I_a$) is equivalent to the effective precipitation producing runoff for a storm event. The initial abstraction was originally set at 20% of S based on calibrations performed in the development of the model. This simplified the CN method to one independent parameter, P, once the site CN was defined.

$$Q = \frac{(P-0.2S)^2}{(P+0.8S)} \quad where\ P \geq 0.2S,\ otherwise\ Q = 0 \qquad (5\text{-}3)$$

Once the CN for a site has been measured by land cover and soils analysis, Equation 5-3 can be used to predict the runoff depth for any given rainfall.

Use of the CN method for runoff prediction applications has been questioned in some areas due to its wide use for a variety of hydrologic conditions that were not considered in its development (Ponce and Hawkins, 1996). Proper model parameter selection is crucial for reliable estimates of direct runoff (DR), but even this may not produce realistic results for some hydrologic conditions due to the assumptions that the model makes in regards to runoff generation. Though the model is simple and generally reliable for watershed outflow prediction in many cases, it often lacks true representation of the physical processes involved in runoff generation (Boughton, 1989). Representation of different hydrologic conditions can be accomplished by varying the parameters of the model to more accurately reflect watershed characteristics and conditions, but this is rarely conducted for the sake of simplicity despite notice in TR-55 of site-specific deviation from typical CN method applications (Ponce and Hawkins, 1996).

Improper selection of site CN can be an initial source of error in using the CN method. DR estimates are more sensitive to the CN than other parameters. For examples, Boughton (1989) has shown that a 15-20% change in CN almost doubles or halves DR predictions. Curve number selection involves the classification of site conditions by discrete categories as defined by TR-55. Natural deviation among these conditions and the potential for misclassification may produce unrealistic runoff estimates due to incorrect CN selection, especially for forested watersheds (Hawkins, 1993). Dual HSG soils further complicate CN selection, as the additional site classification between drained and undrained produces a large difference in runoff estimates. Discrete measurement of site CN is difficult because runoff generation is variable between storm events. The CN has been interpreted as a random variable that ranges for any given storm based on the ARC (Hjelmfelt, 1991; Van Mullem et al., 2002). The CN method offers very little guidance on accounting for differences in runoff generation between dry and wet conditions. ARC was initially based on 5-day antecedent precipitation index (API), but this index was later revised due to differences in regional definitions for site moisture (Ponce and Hawkins, 1996). No specific CN method guidelines for ARC determination are currently offered in TR-55. CN tables for site determination are listed for the average ARC (CN-II), interpreted as the median CN measured by analysis of rainfall and runoff data. A correction must be applied to the CN-II for the dry ARC (CN-I) and wet ARC (CN-III) (Ponce and Hawkins, 1996). These values are considered probabilistic upper and lower limits for runoff generation for a given site based on the range of soil moisture conditions (Hjelmfelt, 1991; Ponce and Hawkins, 1996). Accounting for differences in runoff generation according to ARC is based upon user discretion and subsequent parameter adjustments. Selection of the CN that is best suited to a given watershed can be difficult and prone to error.

The initial abstraction was originally set at 20% of S to simplify the CN method to one parameter. This does not account for differences in site conditions, hydrology, or runoff generation mechanisms. This parameter determines the effective precipitation that contributes to direct runoff generation. Woodward et al. (2003) showed that a level of 5% is more realistic by comparing data from over 300 watersheds over the eastern two-thirds of the United States. Lim et al. (2006) supported this lower measure with GIS modeling in a watershed in the Midwestern United States. Despite this, adoption of a median value for initial abstraction for all CN applications does not reflect regional watershed differences in runoff generation. Ponce and Hawkins (1996) suggested from the work of Bosznay (1989) and Ramasastri and Seth (1985) that initial abstraction might be better interpreted as a regional parameter that reflects differences in geography and climate for better model results. TR-55 notes that care should be taken to ensure that the assumptions made in using this initial abstraction term reflect field measurements. The use of a different value for initial abstraction requires changes in CN tables due to associated changes in CN equations. Widespread use of the CN method is a function of the relative simplicity and reliability of the model, but sources for error in parameter estimation should be considered for model refinements.

OBJECTIVES

Towards the goal of the assessment of the CN method for LCP headwater catchments, storm event rainfall and runoff data were measured to

1. Compare the CN selected using the TR-55 method for two LCP watersheds to back-calculated CNs from storm event data and curve-fitting to determine if there is a difference.
2. Assess the use of $I_a = 0.2S$ by comparison to event-based measurements of I_a derived from storm event hydrographs.
3. Assess the relationship between measures of ARC and back-calculated CNs to define the determining factors for variability in runoff generation for LCP headwater streams.

The goal of this study is not to use event data obtained from 3 years of storm events on two LCP watersheds to define CN model parameters. CN parameter definition typically involves the use of much longer datasets that cover many watersheds (Hawkins, 1993). Rather, this study serves to investigate assumptions made in use of the CN method that may not be applicable to LCP hydrologic conditions and determine if adaptation of the method may be necessary for better modeling results of runoff estimates for headwater catchments in the LCP.

METHODS

Site Descriptions

Rainfall and outflow were monitored from 2008-2011 on two first-order Lower Coastal Plain watersheds. Upper Debidue Creek (UDC) in Bannockburn Plantation (33.38° N, 79.17° W), located in coastal Georgetown County, is a 100 ha freshwater non-tidal watershed that has been slated for development. Potential runoff created by future development is of concern on this tract. Existing downstream development is already experiencing water quantity issues that have forced overflow routing to nearby Waccamaw River to avoid discharges to protected downstream tidal marshes at North Inlet. Watershed 80 (WS80), a tributary of Turkey Creek located in the Francis Marion National Forest (33.15°N 79.8°W), is a 163 ha freshwater non-tidal watershed that is federally protected and serves as an undeveloped reference watershed. The location and monitoring design for each watershed is given in Figure 5.1.

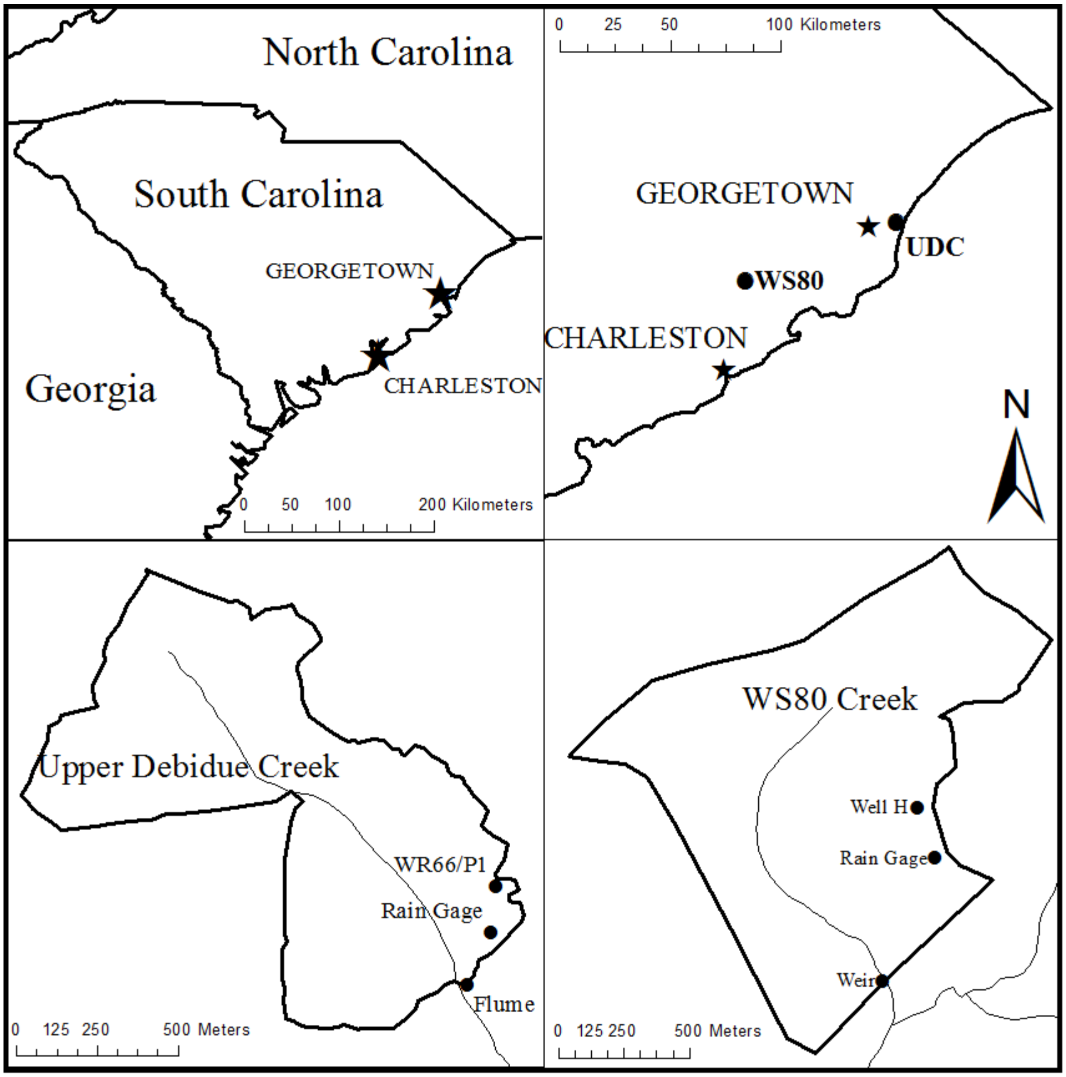

Figure 5.1. Upper Debidue Creek and Watershed 80 locations and monitoring networks.

Both watersheds are typified by low gradient topography and shallow water table conditions. Each landscape is currently dominated by forested wetlands with mixed hardwood lowlands and upland pine stands. The primary soils in the UDC watershed are Lynn Haven (HSG B/D) and Leon (HSG A/D). These soils are formed of sandy marine sediment, are associated with very low gradient conditions, are highly permeable, and poorly drained (USDA 1980). The primary soils at WS80 are Wahee (HSG D), Meggett (HSG D), Craven (HSG C), and Bethera (HSG D). These soils are formed of clayey Coastal Plain sediments and are typical of areas with low gradient topography (USDA, 1974). These soils are poorly drained with high available water content and have lower permeability than sandy soils. The two watersheds are 75 km apart and in a humid subtropical climatic zone characterized by short mild winters and long hot summers (Sun et al., 2000). Soil composition and land cover data for both watersheds were obtained from SC DNR to perform watershed CN calculations. All areas were classified as "woods in good condition" per the TR-55 guidance (USDA, 1986).

Data Collection

Rainfall, stream outflow, and groundwater elevation data were collected from 2008-2011 (Fig. 5.1). Tipping bucket rain gages (Onset® Hobo™, Bourne, MA at UDC and WS80) located in both watersheds were used to quantify local hourly rainfall totals. Groundwater elevations were monitored at upland locations near the watershed boundary on both UDC and WS80. At UDC, a 3- m deep water table well with a pressure transducer was located in an upland pine area near the watershed boundary (WR66, Fig. 5.1). The pressure transducer was replaced with a Solinst™ levelogger® in March 2011. Groundwater elevation was monitored in the upland area at WS80 (Well H, Fig. 5.1) by a WL16 logger® (Global Water, Gold River, CA). Watershed outflow at UDC was estimated using a 0.6 m modified Parshall flume located immediately downstream of a road culvert. Outflows were corrected for submergence in the flume according to equations developed by Peck (1998). A threshold of 0.85 for submergence was set for measured outflow in accordance with the correction equations. Additional instrumentation details for UDC are provided by Hitchcock et al. (2009). At WS80, watershed outflow was estimated by measuring stage over a compound weir. Additional instrumentation details for flow measurement in WS80 are provided by Harder et al. (2007).

Data Assessment

Storm events were selected from data for 2008-2011 at both watersheds in order to assess the rainfall response over comparable local climate and moisture conditions for the same time period because of the proximity of the two sites. Storms were selected given that they met the following criteria: (1) rainfall greater than 20 mm, (2) outflow was generated, and (3) event displayed a simple hydrograph with a recession limb of sufficient length to perform graphical analysis. These criteria allowed for selection of a sufficient number of storms from the three-year period at both watersheds that ranged from substantial, higher frequency storm events up to larger storms of much less frequency.

DR estimates for storm events were calculated by the hydrograph separation method developed by Williams (2007) and discussed in length in Chapter 4. CN back-calculation was performed by solving the CN equation by the quadratic formula for S, and the negative root was taken as the solution in order to preserve the relationship that $P = Q$ when $S = 0$ (Hawkins, 1993). Values of S were back-calculated by

$$S = 5 * [P + 2Q - \sqrt{4Q^2 + 5PQ}] \quad (5\text{-}4)$$

and these were converted to respective CNs by Equation 5-1. Back-calculated CNs were determined using $I_a = 0.2*S$ in order to assess measured data against parameters typically used for TR-55 applications.

Ordered pairs of (P, Q) data were used to perform non-linear curve-fitting to the CN Equation 5-3 to determine the optimal CN from rainfall and direct runoff estimates by least squares analysis. Software employing the generalized reduced gradient algorithm (Fylstra et al., 1998) was used to obtain a parameter estimate for S which was converted to CN by Equation 5-1. The inequality P > 0.2S was converted to S < 5P and used as a parameter constraint. The lowest observed value of P for the storm events was substituted into the constraint so that results would satisfy Equation 5-3. The coefficient of determination (COD) for curve-fitting was calculated as

$$COD = \frac{SS_{Total} - SS_{Err}}{SS_{Total}} \quad (5\text{-}5)$$

The sum of the squares of the error, SS_{Err}, is a measure of the residual between observed data and the best-fit curve and is calculated by

$$SS_{Err} = \sum_{i=1}^{i} (y_i - \hat{y}) \quad (5\text{-}6)$$

where i is the number of observations, y_i is the i^{th} observation, and $\hat{y}$ is the i^{th} model estimate. The total sum of squares, SS_{Total}, represents the residual between observed data and its mean value and is calculated by

$$SS_{Total} = \sum_{i=1}^{i} (y_i - \bar{y}) \quad (5\text{-}7)$$

where $\bar{y}$ is the average of the observations. As the COD approaches a value of 1.0, curve-fit results account for a greater portion of the variability in the overall data set.

Negative values for the COD are possible for non-linear regression and they indicate that the mean for observed data is a better predictor than the best-fit curve. Statistical inferences for results of non-linear regression are difficult to compare between non-linear models and between data sets due to compounded uncertainty involved in the regression process (Motulsky and Rasnas, 1987). The COD values from non-linear regression will only be used in order to discuss differences between curve-fitting results and TR-55 expectations for runoff generation. Sample means for back-calculated CNs were compared first by an F-test to determine if the variances were equal. Depending on this outcome, the appropriate two-sample t-test for equal or unequal variance was used to determine if there was a difference between the means.

Initial abstraction was measured for each storm event as the amount of rainfall that preceded the initial rise in the hydrograph. I_a estimates were divided by values for S back-calculated from storm event (P, Q) pairs. The I_a/S ratios for all storm events on each watershed were assessed to determine median values for comparison to the accepted 20% value currently used. The 5-day API, measured as rainfall that occurred during the 120-hour period prior to a given storm event, and antecedent water table elevation were compiled for each storm as estimates of ARC for comparison to runoff generation. Linear regression was performed to assess the relationship between these estimates of ARC and back-calculated CNs at each watershed.

RESULTS AND DISCUSSION

Hydrograph separation was performed for 23 storm events at UDC and 20 storm events at WS80 that met selection criteria as previously described. These storm events ranged in size from 20 – 87 mm at UDC and 29 – 154 mm at WS80. A histogram of storm event size for each watershed is provided in Figure 5.2. Back-calculated CNs ranged from 46 to 90 at UDC (Table 5.1) and from 42 to 89 at WS80 (Table 5.2).

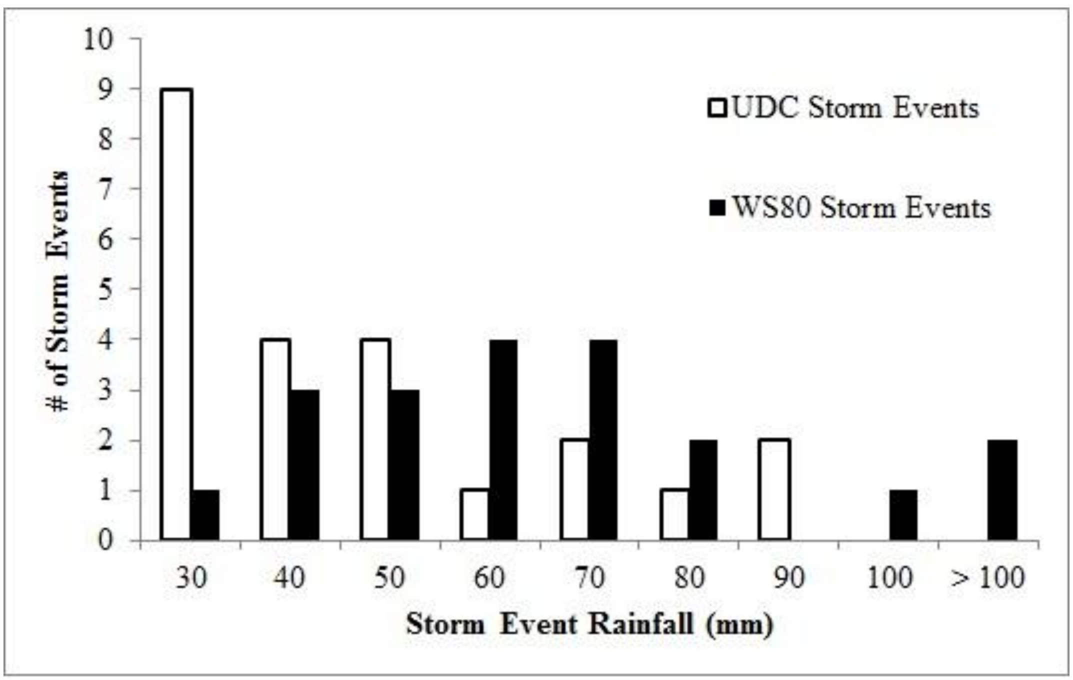

Figure 5.2. Event-based histogram for Upper Debidue Creek and Watershed 80 storm events.

The large number of storm events smaller than 30 mm at UDC has been taken into consideration in the analysis and will be discussed throughout where it applies. Storm event rainfall, DR, back-calculated CN, and measures of ARC have been summarized for UDC storm events in Table 5.1 and WS80 storm events in Table 5.2.

Table 5.1. Summary of storm events for Upper Debidue Creek, back-calculated Curve Numbers, and Antecedent Runoff Condition measures.

Date	**Rainfall (mm)**	**Direct Runoff (mm)**	**Curve Number**	**Initial Outflow (m^3/hr)**	**Water Table Elev. (m asl)**	**5-day API (mm)**
7/24/2008	30	1	70	17	NA	11
9/5/2008	87	12	59	42	NA	0
9/11/2008	25	4	86	166	NA	81
9/16/2008	47	12	80	159	NA	0
9/25/2008	42	10	80	134	NA	1
3/1/2009	40	3	72	16	3.19	1
4/2/2009	60	9	70	83	3.38	18
8/28/2009	68	0	46	0	2.34	0
11/10/2009	78	2	49	0	2.42	5
1/16/2010	22	2	83	119	3.60	0
1/25/2010	23	5	89	253	3.73	17
2/2/2010	27	7	87	311	3.74	20
3/2/2010	24	8	90	158	3.61	0
5/4/2010	23	0	73	2	3.25	0
6/20/2010	36	0	60	0	2.85	0
6/30/2010	35	0	63	0	2.93	36
7/10/2010	35	0	65	0	2.95	19
8/1/2010	24	0	72	0	3.03	6
8/13/2010	40	1	66	23	2.97	40
8/19/2010	25	0	72	2	3.29	36
6/29/2011	20	0	76	0	2.49	0
8/6/2011	81	3	51	24	2.66	6
8/25/2011	67	5	59	51	2.76	13

Table 5.2. Summary of storm events for Watershed 80, back-calculated Curve Numbers, and Antecedent Runoff Condition measures.

Date	Rainfall (mm)	Direct Runoff (mm)	Curve Number	Initial Outflow (m^3/hr)	Water Table Elev. (m asl)	5-day API (mm)
8/21/2008	37	1	66	14	8.94	31
9/5/2008	98	2	42	0	8.26	1
9/9/2008	113	31	64	31	8.97	98
9/25/2008	65	5	61	0	8.62	1
10/24/2008	154	88	76	2	8.79	0
11/29/2008	47	4	68	2	8.64	0
3/1/2009	58	14	75	6	8.80	4
4/2/2009	67	23	78	49	9.02	35
7/16/2009	41	1	62	1	8.69	30
7/22/2009	29	0	68	1	8.80	0
8/31/2009	57	1	54	0	7.85	48
11/11/2009	70	0	44	0	7.53	2
12/18/2009	67	26	80	40	9.06	12
12/25/2009	31	7	84	77	9.06	3
1/16/2010	51	9	74	13	8.97	0
1/25/2010	42	19	89	167	9.08	25
3/28/2010	31	4	81	14	8.97	0
5/4/2010	52	0	52	0	8.07	0
9/29/2010	75	12	64	102	9.09	147
2/2/2011	66	10	67	4	8.88	12

Curve Number Estimates

CN calculations using TR-55 methods of selection by soils and land cover analysis for UDC determined a watershed CN of 75 in the undrained condition and a CN of 37 if adequately drained. A CN of 75 was calculated for WS80 by this method and it does not have any dual HSG soils. The National Engineering Handbook recommends the use of the drained HSG for CN determination only if the seasonal high water table remains below 0.6 m below ground surface (USDA, 2007). Water table elevations above this level are not typically observed at UDC at the groundwater well near the watershed boundary (WR66) but in the more saturated riparian zones, the water table is typically closer to the surface for wet ARC. Variable water table elevation makes discrete classification between drained and undrained for these dual HSG soils difficult. The interpretation of the CN as a random variable supports this range of potential CNs. It is evident that a single CN for watersheds containing these dual HSG soils does not adequately describe runoff generation due to variable water table conditions.

Storm event rainfall and direct runoff were plotted as (P, Q) pairs. Curves representing the TR-55 determined CN values as defined by Equation 5-3 were plotted with them for comparison in Figures 5.3 and 5.4.

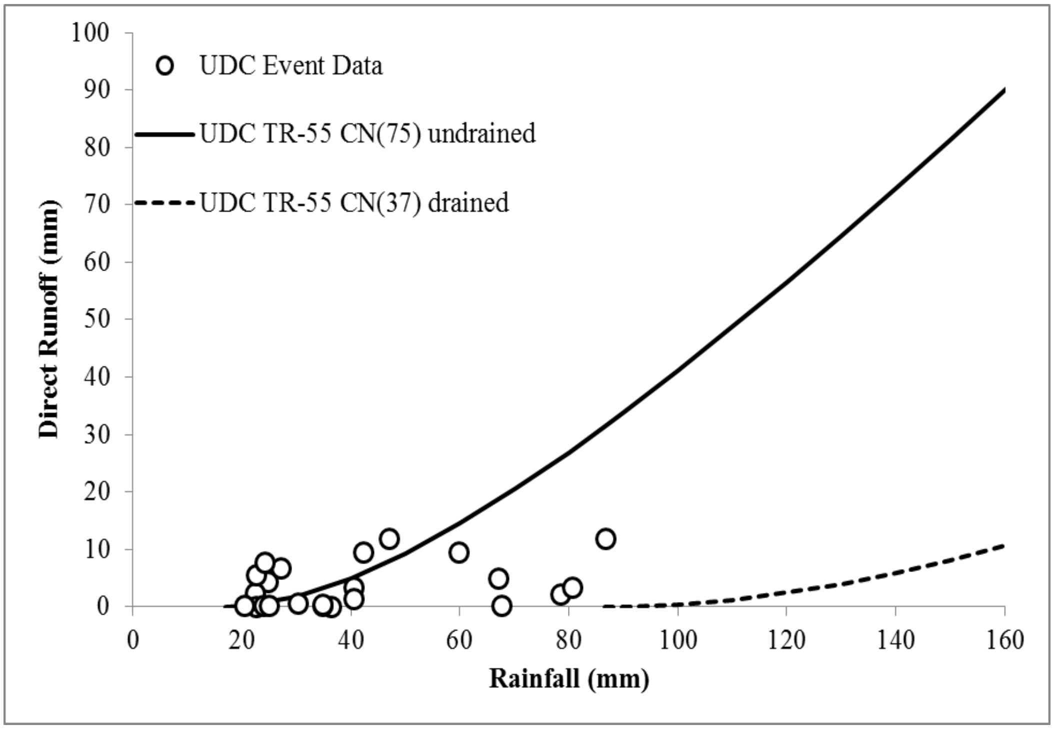

Figure 5.3. Upper Debidue Creek storm events plotted with curves for CN=75 and CN=37.

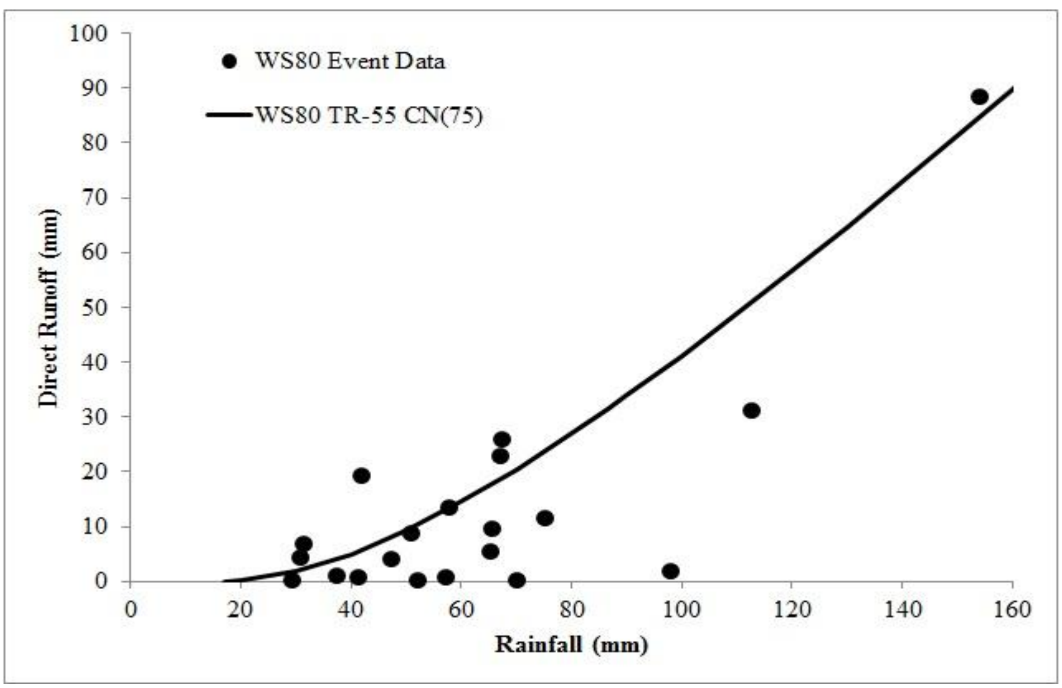

Figure 5.4. Watershed 80 storm events plotted with curve for CN=75.

The COD for the CN=75 curve was calculated at each watershed to assess the measured data against TR-55 CN selection. The undrained CN for UDC was not included in this analysis because only one storm event had sufficient rainfall ($P > 86$ mm) to have produced direct runoff for a CN of 37. All storm events had back-calculated CNs higher than 37 and this does not serve as a good median CN value for UDC. COD values were calculated by Equation 5-5 between the curve for CN=75 and measured data. At UDC, the COD for the data equaled (-5.5) and can be interpreted as a poor fit to the curve. The negative value indicates that event data at UDC would be better modeled by a straight line at the mean of measured DR. This relationship seems to be a function of the scatter of the data and the range of DR values measured for similar rainfall depths. At WS80, the COD was 0.59, indicating that the curve for CN=75 is a relatively good fit for the data. TR-55 CN values are intended to represent average conditions and have been interpreted as the median CN for a site, splitting measured event data sets in half (Hjlemfelt, 1991). Visual inspection of the data in relation to the curve for CN=75 indicates that a curve with a lower CN would better split the measured data in half for both watersheds (CN values for (P, Q) data decrease towards the lower right-hand side of the plot). Descriptive statistics for back-calculated CNs are summarized in Table 5.3.

Table 5.3. Select descriptive statistics for storm event back-calculated Curve Numbers at Upper Debidue Creek and Watershed 80.

	UDC	**WS80**
Count (n)	23	20
Mean	70.37	67.51
Std. Error	2.65	2.88
Median	71.51	67.56
Std. Dev.	12.72	12.87
Minimum	45.82	41.69
Maximum	90.50	89.06

Median values for back-calculated CN were lower than 75 for each watershed with 72 at UDC and 68 at WS80. The mean CN for storm events at UDC was 70 and the mean was found to be statistically less than 75 ($\alpha = 0.05$, $p = 0.047$). The mean CN for storm events at WS80 was 68, also statistically lower than 75 ($\alpha = 0.05$, $p < 0.01$). This evidence suggests that DR estimates for the storm events on these two watersheds did not produce as much runoff on average as TR-55 CN estimates would have predicted.

Non-linear regression was performed by the least squares method to determine the best-fit CN curves for measured data at UDC and WS80. Because the least-squares method minimizes the SS_{Err}, or the sum of the squares of the residuals between measured data and the best-fit curve, solutions by this method should approximately split the data in half and serve as a good approximation of the median CN curve. Results are depicted graphically in Figures 5.5 and 5.6 and summarized in Table 5.4.

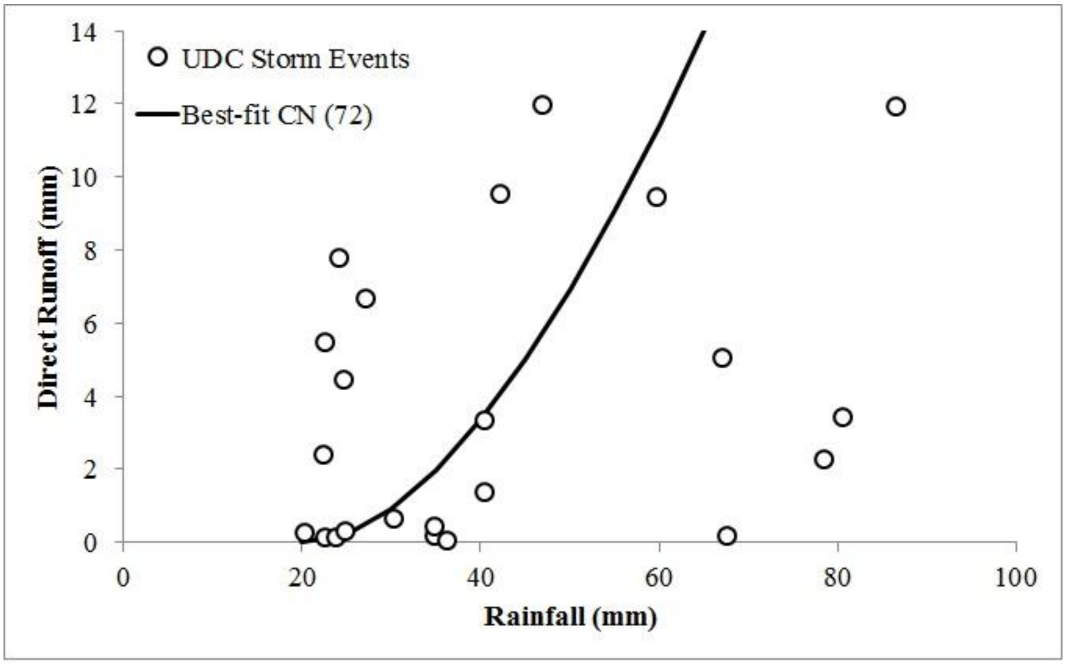

Figure 5.5. Best-fit Curve Number curve for Upper Debidue Creek storm event rainfall and direct runoff.

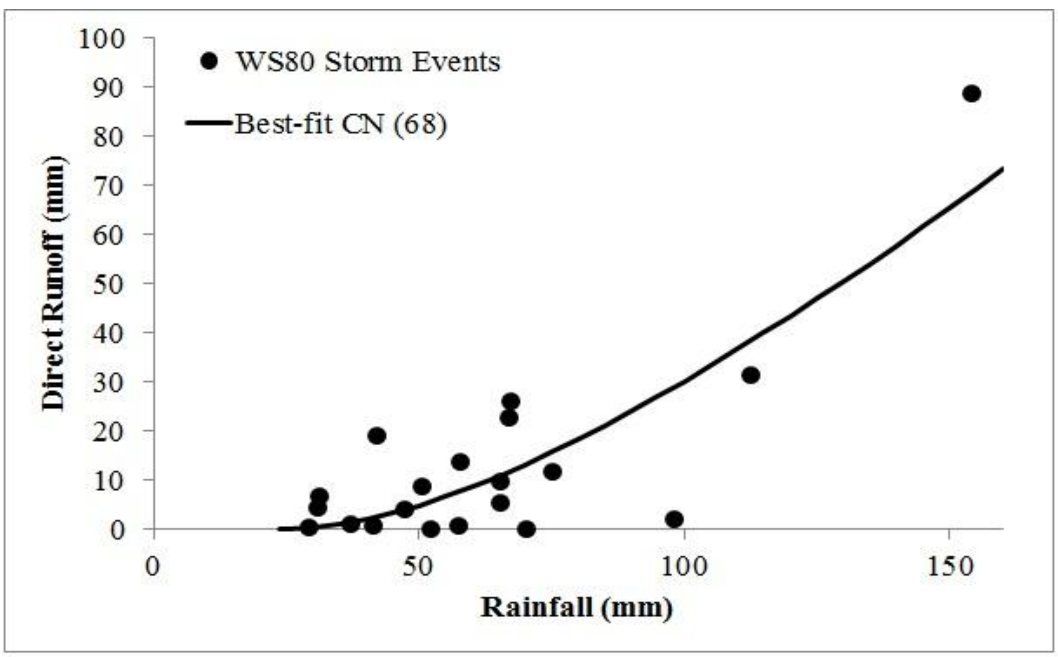

Figure 5.6. Best-fit Curve Number curve for Watershed 80 storm event rainfall and direct runoff.

Table 5.4. Summary of non-linear regression by least squares of CN curves for UDC and WS80.

	UDC	**WS80**
S	3.94	4.65
CN	**72**	**68**
SS_{Err}	1,531	2,169
SS_{Total}	351	7,667
R^2	-3.37	0.72

Results of the best-fit curves were lower than 75 at both watersheds with a CN of 72 calculated for UDC and a CN of 68 calculated for WS80. These results are equivalent to the median values of back-calculated CNs measured for all storm events. The constraint ($S < 5P$) was binding for the UDC regression, and the resulting best-fit curve does not split the data evenly. In previous studies, the influence of smaller storms on CN measurement from event data has been shown to have a positive bias on CN estimates (Hawkins et al., 1985; Hjelmfelt, 1991). Removal of this constraint would result in a lower best-fit CN, but storm events with lower rainfall would have to be eliminated from the data set in order to fit only the right side of the quadratic formed by Equation 5-3. Least squares results measured for storm events at each watershed show improvement of the COD compared to that measured for the CN=75 curve that represents the TR-55 derived watershed CN. At WS80 the COD value increased from 0.59 to 0.72 for curve-fitting results. This analysis indicates that the best-fit CN of 68 accounts for a larger portion of the variability in storm event data than the curve derived using the TR-55 CN of 75. At UDC the COD value increased from (-5.5) to (-3.4). While this does indicate that the best-fit CN of 72 accounts for more of the variability in the storm event data, the data is still better modeled by the mean DR observed. The storm event data does not fit well to the CN curve equation as displayed by negative results for COD. The range of DR generation that is observed for similar rainfall amounts is not described well by a single curve. While this may be a function of observed storm events, it is possible that the storm event data might best be modeled by two separate CNs that are related to ARC and water table position. Dual HSG soils may range between lower runoff generation characteristic of the drained condition and higher runoff generation characteristic of the undrained condition solely based on ARC and water table position that is not related to physical drainage methods. In this study, an assessment of runoff generation based on variables that characterize the ARC for these watersheds will be performed.

Initial Abstraction

Descriptive statistics for the ratio I_a/S obtained from measured I_a and back-calculated S are summarized in Table 5.5.

Table 5.5. Summary of descriptive statistics for Ia/S ratios at Upper Debidue Creek and Watershed 80.

	UDC	**WS80**
Count (n)	23	20
Mean	0.30	0.13
Standard Error	0.26	0.05
Median	0.01	0.07
Standard Deviation	1.25	0.22
Minimum	0.00	0.00
Maximum	5.99	0.84

Results are consistent with observations made by Woodward et al. (2003) that I_a/S ratios range from storm to storm on a watershed but that measured values are closer to 5% of S than the 20% that is recommended in TR-55. Based on Woodward et al. (2003), the median value for event analysis of I_a/S ratios was taken as the value for a watershed. This was 0.01 at UDC and 0.07 at WS80. These results agree with assertions that 5% of S may be a better ratio for CN runoff estimates (Lim et al., 2006; Ponce and Hawkins, 1996; Woodward et al., 2003). Concerns over the effects of storm size on results were addressed by Woodward et al. (2003) by using only storm events for which $(P- I_a) > 25.4$ mm. Descriptive statistics for I_a/S ratios that included only storm events with $(P-I_a) > 25.4$ are summarized in Table 5.6. The elimination of smaller storms altered the sample mean, but the median values for I_a/S were consistent with results over all storm events.

Table 5.6. Summary of descriptive statistics for Ia/S ratios of storm events satisfying (P-I_a) > 25.4 mm.

	UDC	WS80
Count (n)	10	15
Mean	0.02	0.14
Standard Error	0.01	0.07
Median	0.01	0.03
Standard Deviation	0.02	0.26
Minimum	0.00	0.00
Maximum	0.05	0.84

Runoff Generation and ARC

Previous classification of CN for dry and wet ARC were applied based on the 5-day API. There is a weak relationship (α= 0.05) between 5-day API and back-calculated CN for storm events at UDC (p = 0.38) or WS80 (p = 0.79) (Figs. 5.7 and 5.8).

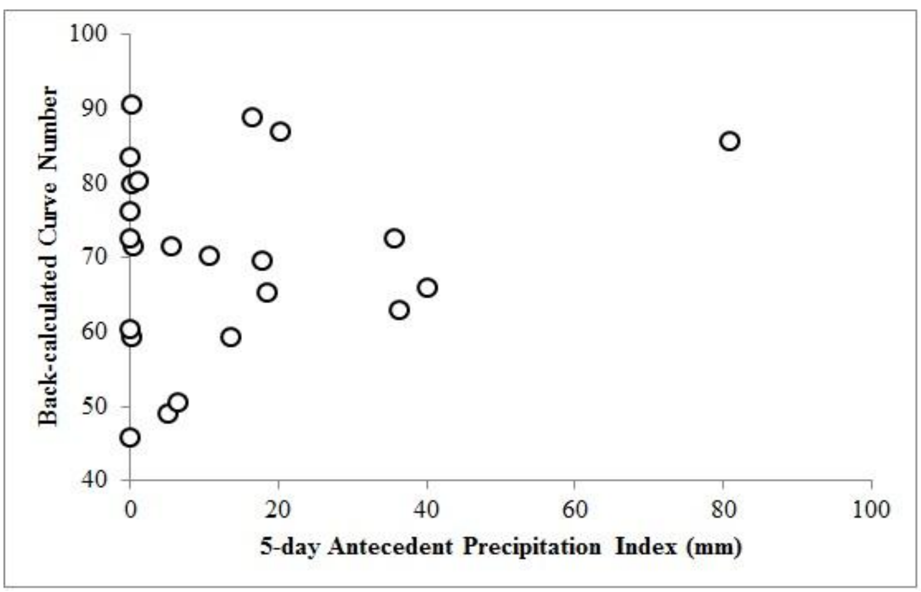

Figure 5.7. Back-calculated Curve Numbers plotted against 5-day Antecedent Precipitation Index for Upper Debidue Creek storm events.

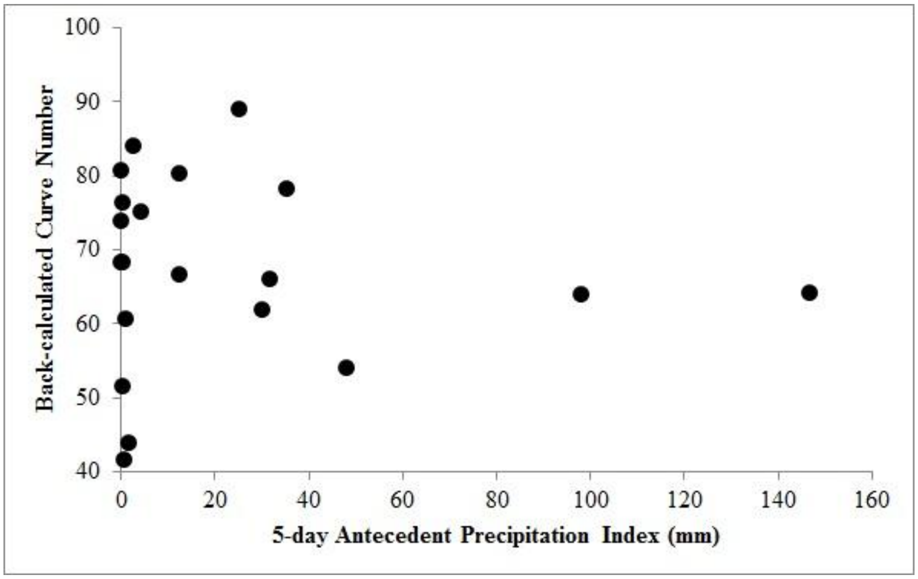

Figure 5.8. Back-calculated Curve Numbers plotted against 5-day Antecedent Precipitation Index for Watershed 80 storm events.

A range of runoff generation measured by back-calculated CNs was observed for 5-day API measurements and no significant relationship was measured by linear regression. 5-day API is not a good measure of ARC for headwater catchments in the LCP. It was hypothesized that the range in runoff generation on LCP headwater catchments is related to antecedent water table elevation. This hypothesis is supported when the relationship between antecedent water table elevation and back-calculated CN is observed (Figs. 5.9 and 5.10). Previously, TR-55 had accounted for a range in storm event CN according to the antecedent precipitation.

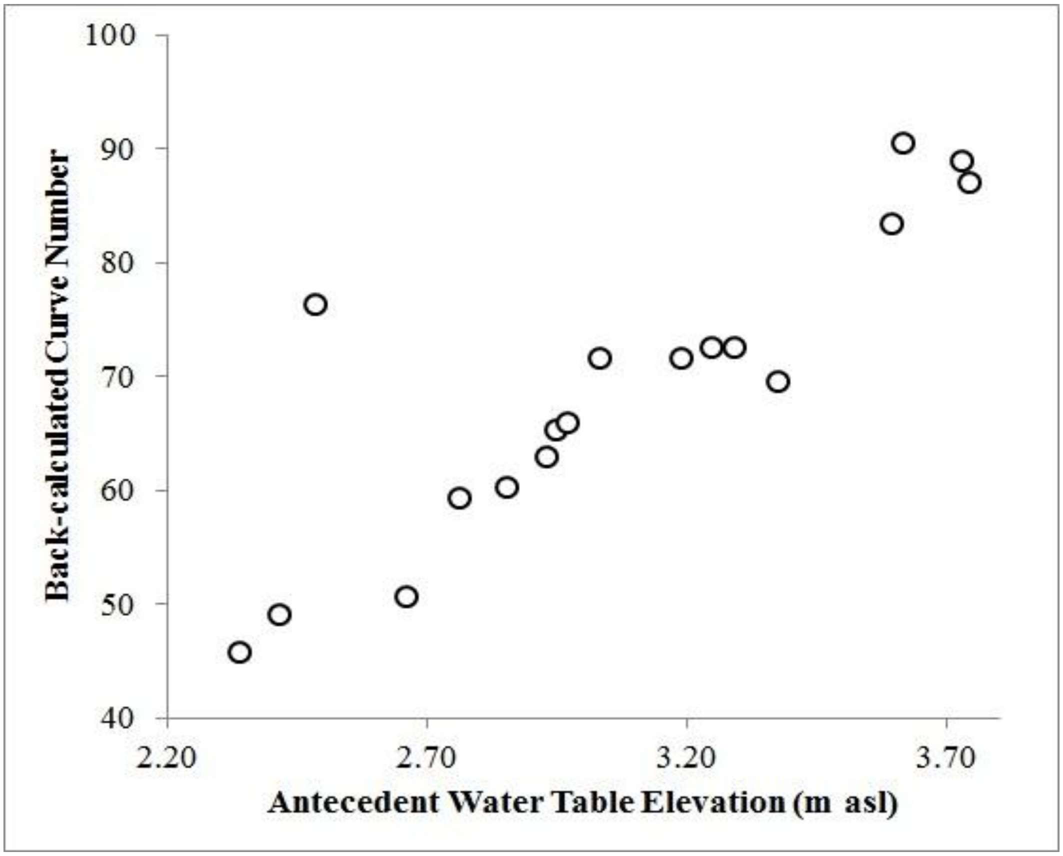

Figure 5.9. Back-calculated Curve Numbers plotted against antecedent water table elevation for Upper Debidue Creek storm events.

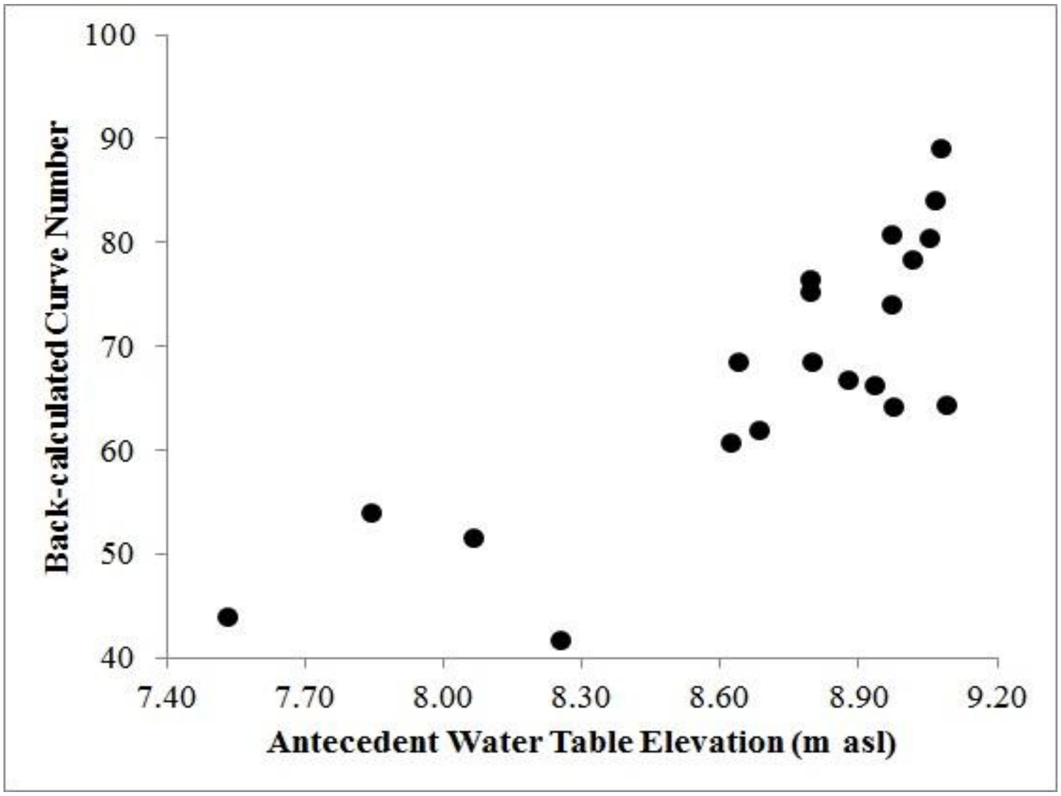

Figure 5.10. Back-calculated Curve Numbers plotted against antecedent water table elevation for Watershed 80 storm events.

The relationship between antecedent water table elevation and the back-calculated CN is strong and indicates that water table elevations determine runoff generation in these LCP headwater streams. Results of linear regression illustrate this close relationship at UDC ($r^2 = 0.75$, $p < 0.01$) and at WS80 ($r^2 = 0.66$, $p < 0.01$). Lower water table elevations are associated with lower CNs. This relationship indicates that DR generation is lower when water table elevations are lower and these conditions define dry ARC. Back-calculated CNs increase along with water table elevation, and runoff generation is higher when the water table elevation is closer to the ground surface indicative of wet ARC. These results indicate that runoff generation varies between storm events based upon the position of the water table on these watersheds. The determination of ARC is best performed based upon antecedent water table elevation because of the strength of this relationship.

Variation in runoff generation as a function of water table elevation on LCP headwater streams must also follow seasonal trends in groundwater elevation. These water table elevation trends have been observed to be a function of seasonal differences in ET. Higher water table elevation and wet conditions dominate during the winter months when ET rates are low. Lower water table elevation and dry conditions dominate during the summer months when ET rates are high. This trend is shown for both watersheds over a year of continuous groundwater elevation observations (Figs. 5.11 and 5.12).

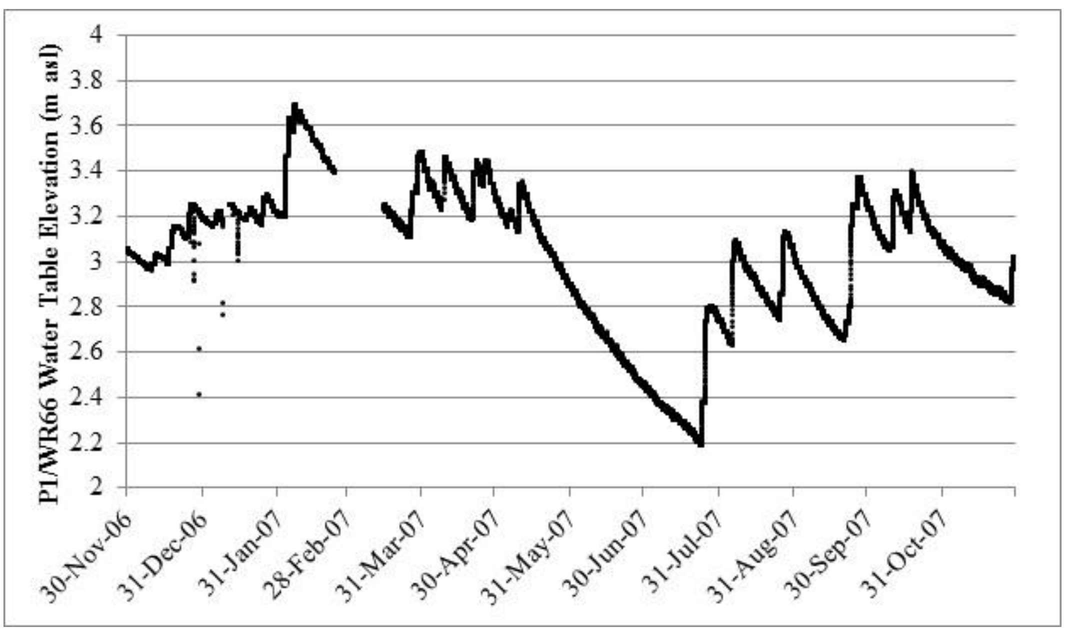

Figure 5.11. Groundwater elevation at P1/WR66 on Upper Debidue Creek for Dec. 1, 2010 – Nov. 30, 2011.

Figure 5.12. Groundwater elevation at Well H at Watershed 80 for Sep. 1, 2010 – Aug. 31, 2011.

Antecedent water table elevations for UDC storm events ranged from 2.34 – 3.74 with a mean elevation of 3.07 m asl. Trends in water table elevation at UDC displayed in Fig. 5.12 show that the water table is above this mean elevation from the end of December 2010 until nearly mid-May 2011. Summer months are dominated by water table elevations lower than the mean elevation of 3.07 m asl, and the water table fluctuates above and below this mean level during the fall months of 2011. Because of the close relationship between water table elevation and DR generation that was observed, above average runoff generation should be expected when the water table is higher than the mean elevation and below average runoff generation would coincide with water table elevations lower than the mean value. The mean value is not statistically significant here and only serves as a reference point for the discussion of water table trends about an "average" condition that would coincide with average ARC and CN-II.

Antecedent water table elevation ranged from 7.53 – 9.09 m asl for WS80 storm events with a mean elevation of 8.70 m asl. Water table elevations over one year displayed in Fig. 5.13 for WS80 show that the water table elevation is near or above this mean level of 8.70 m asl for most of the period from January 2011 through the end of April 2011. Conditions are wet during winter and spring months. Water table elevation falls below this mean level during the summer months when ET rates are high. Sharp rises above the mean level in response to rainfall events during summer months are associated with flashy peaks in groundwater elevation, but wet conditions do not persist because relatively higher ET rates lower the water table elevation rapidly.

Seasonal groundwater trends have implications related to ARC and CN predictions of direct runoff (DR). High water table conditions that characterize wet ARC are dominant during the winter months. DR estimates using the average ARC will be lower than actual runoff generation for high water table conditions and wet ARC, while the opposite is true during the summer months when dry ARC persist. DR estimates using the average ARC will likely be higher than actual runoff generation for low water table elevations and dry conditions. Though the average ARC represents the central tendency for runoff generation for the watersheds of the LCP, runoff generation that is expected by this median CN may be less likely to occur for any given storm event. Wet and dry conditions persist during winter and summer months, respectively, and influence the fluctuation between periods of higher and lower runoff generation related to water table position. The seasonal trend of water table elevation influences ARC on a seasonal basis because of the close relationship observed. This trend in runoff generation should be accounted for in CN applications for LCP headwater catchments. Using the CN for average ARC in all design applications will likely result in systematic errors for DR estimates that are related to seasonal trends in water table elevation and ARC.

CONCLUSIONS

Analysis of rainfall and direct runoff estimates for storm events on two LCP headwater catchments indicate that the CN method does not model the rainfall response for these watersheds very well. The CNs assigned to these watersheds by soil and land cover analyses are higher than estimates that were calculated by analysis of observed data. Dual HSG soils are linked to shallow water table conditions, and runoff generation for these soils may not be modeled well by a single CN. The initial abstraction was measured to be much closer to 5% of S, supporting work that has questioned the TR-55 definition that was set at 20% of S. A strong relationship between antecedent water table elevation and runoff generation as measured by back-calculated CN was observed, and through this work it is recommended that the ARC should be defined in relation to variable water table conditions on LCP headwater catchments. Because seasonal trends in ET influence seasonal trends in water table elevation, the use of the average ARC CN-II for design work could result in systematic error in runoff prediction that under predicts runoff during the winter months and over predicts runoff during the summer months. The range in runoff generation from dry to wet ARC for these watersheds should be accounted for because of their relationship to seasonal variation. Consideration of these departures from accepted TR-55 CN methodology will lead to more accurate runoff predictions for LCP headwater streams.

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CHAPTER SIX

CONCLUSIONS

Toward the goal of the assessment of the rainfall response on forested headwater catchments in the Lower Coastal Plain (LCP) of South Carolina, rainfall, streamflow, and groundwater elevation were monitored on two similar streams. Spatial rainfall variability was assessed at UDC and it was concluded that a single gage was sufficient to accurately measure rainfall for this watershed. Throughfall measurements at UDC did not appear to accurately measure the spatial variation of subcanopy rainfall, but they do indicate a seasonal difference that may influence seasonal trends in streamflow. This seasonal difference may be related to trends in evapotranspiration and the long growing season, and this may warrant further study in the LCP for inclusion in seasonal water budgets. The rainfall response on the two watersheds was measured as total storm flow and direct runoff components of watershed outflows. Variability in runoff generation at UDC and WS80 was related to seasonal trends of evapotranspiration that determine soil moisture conditions and are related to seasonal fluctuations in groundwater elevation. Break point water table elevations were determined for each watershed above which runoff generation was observed to increase sharply. This groundwater elevation may have a physical significance in runoff generation mechanisms. Runoff generation in LCP headwater catchments appears to differ for sandy and clayey soil types. Measures of total storm flow may be more applicable for watersheds with clayey soils and direct runoff determined by graphical hydrograph separation may model runoff generation for watersheds with sandy soils better. This is likely due to differences in the influence of the groundwater aquifer and baseflows on streamflows and runoff generation observed for the two watersheds. The SCS Curve Number method for runoff modeling was compared to measured rainfall and runoff for storm events on both watersheds. Parameter selection by the accepted methodology does not appear to accurately model runoff generation on these LCP headwater catchments. The strong relationship between groundwater elevation and runoff generation should be considered for applications of the Curve Number method in similar watersheds. The effect of seasonal trends in groundwater elevation on the rainfall response for similar streams in the LCP may not be well modeled by the median measure for runoff generation that is typically used due to fluctuating moisture conditions that have seasonal patterns. This work has demonstrated that seasonal trends in soil moisture that are related to low topography and the role of vegetation in these forest watersheds influence groundwater elevations which in turn determine runoff generation and storm flow levels in the rainfall response.

APPENDICES

Appendix A

Storm Event Hydrograph Separations for Upper Debidue Creek

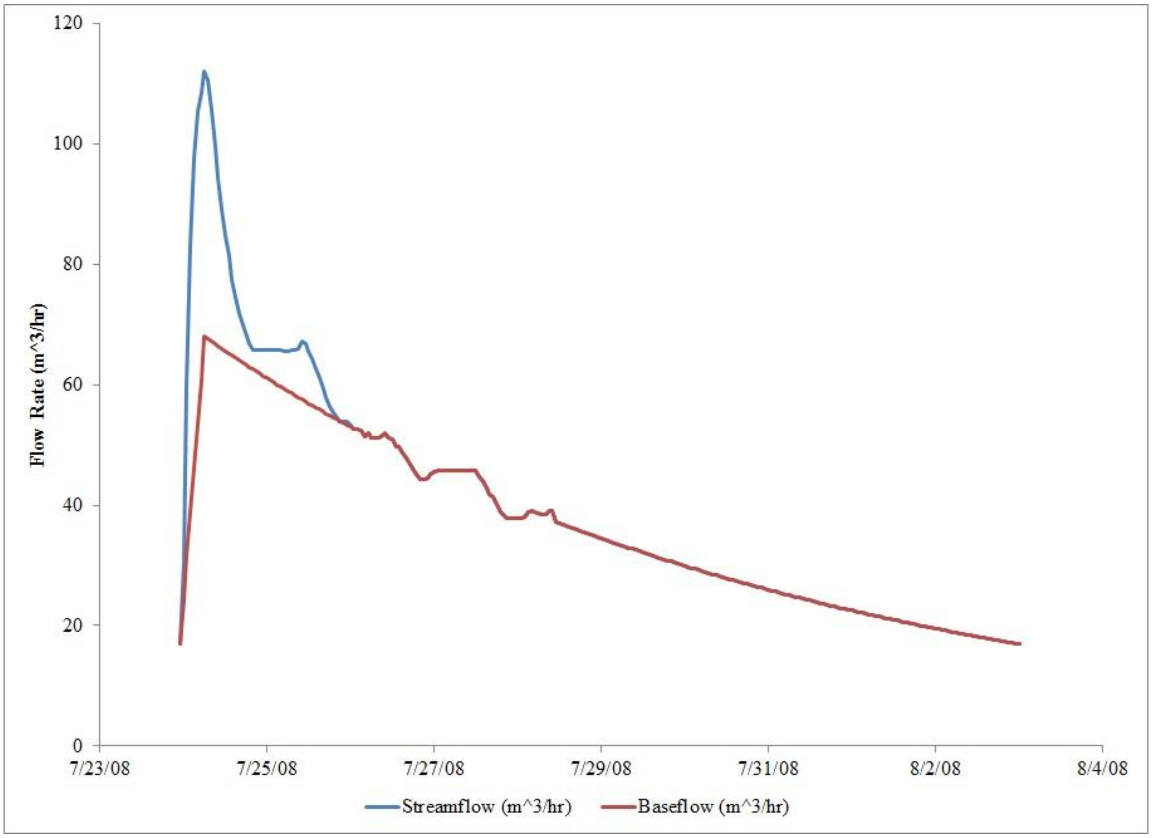


Figure A-1: Hydrograph separation for storm event on 7/24/08.

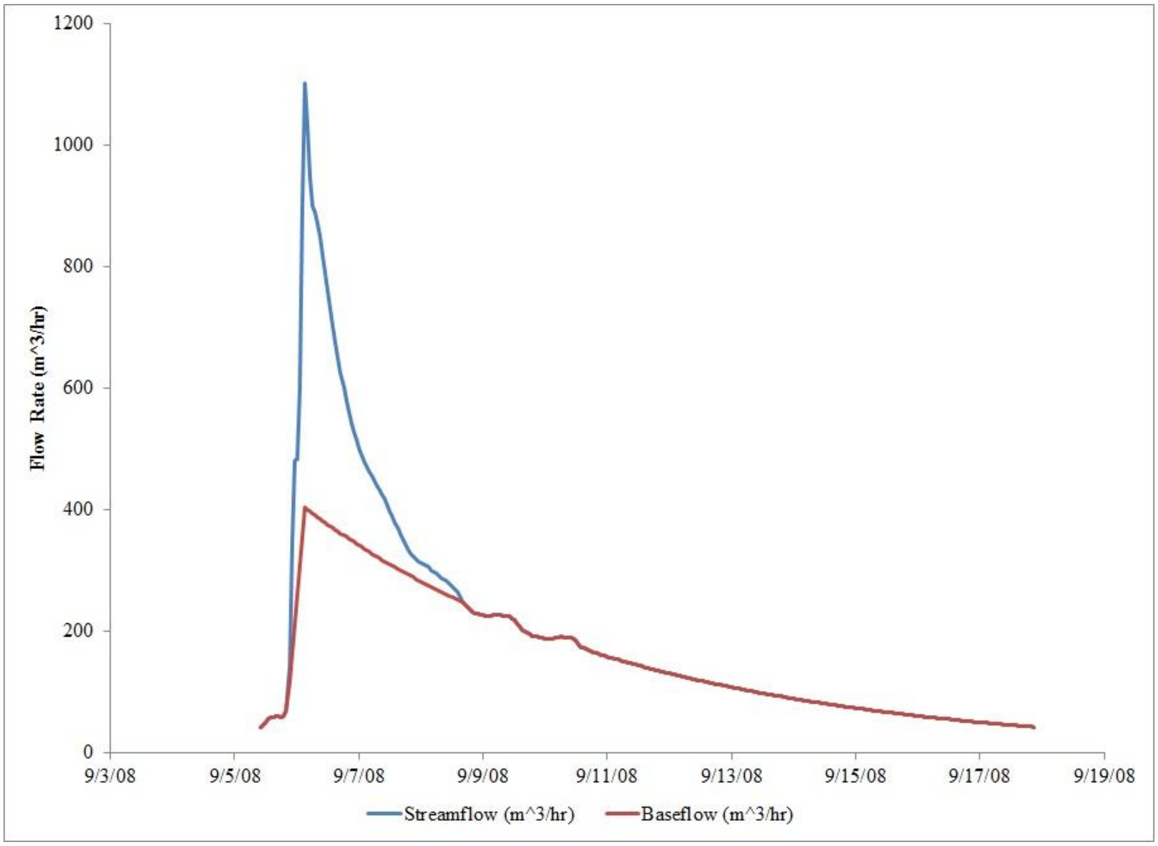


Figure A-2: Hydrograph separation for storm event on 9/5/08.

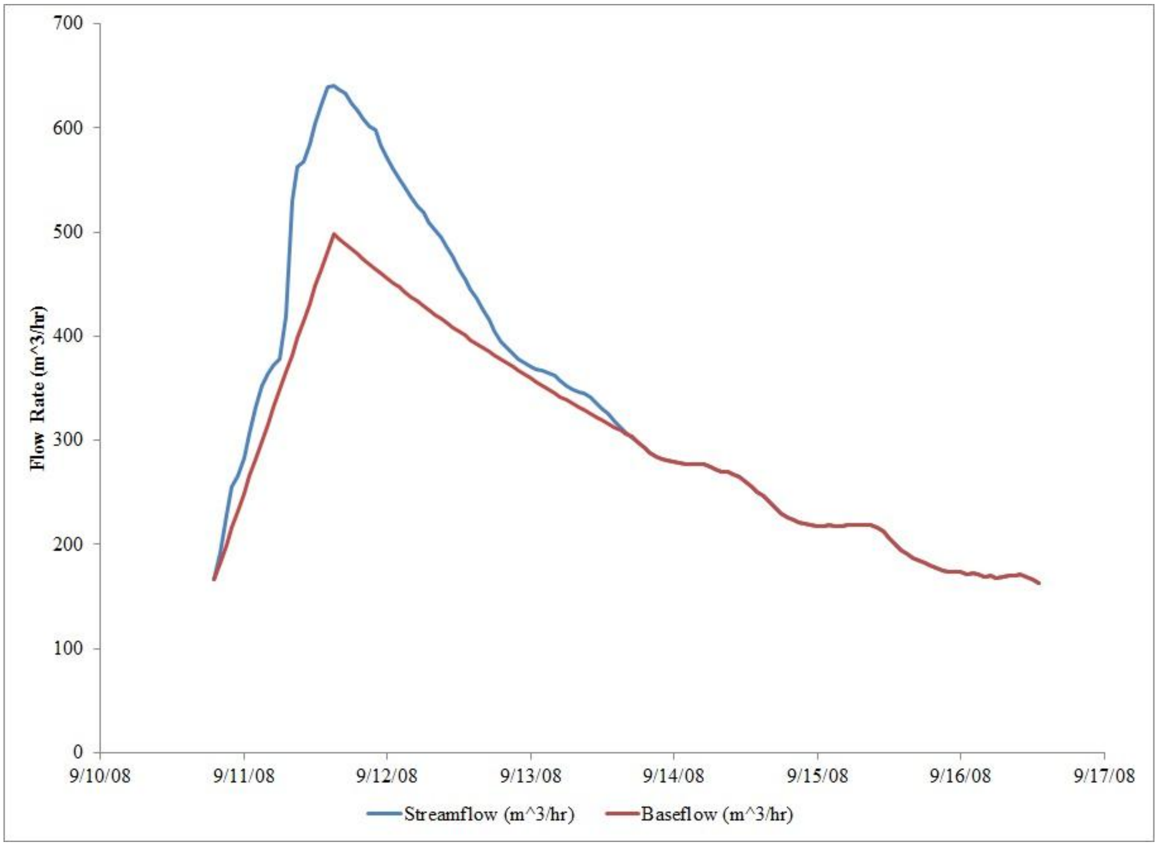


Figure A-3: Hydrograph separation for storm event on 9/11/08.

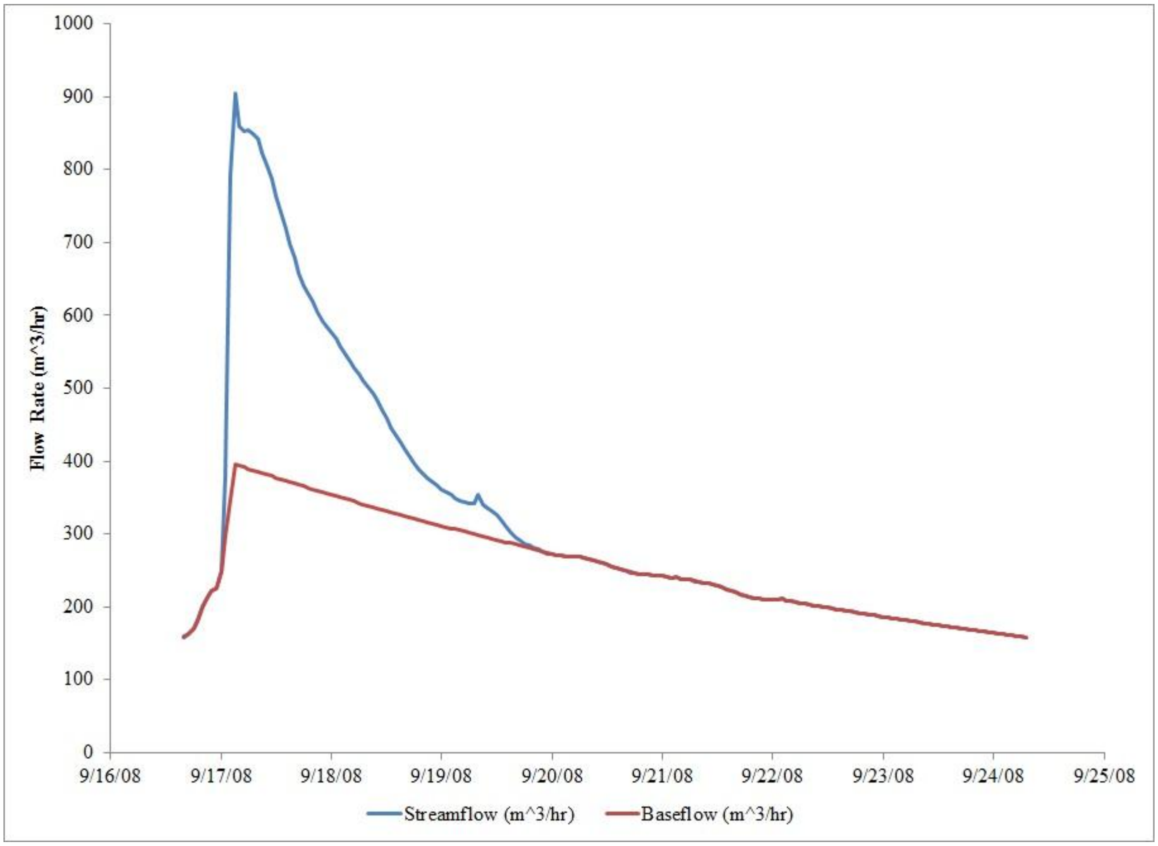


Figure A-4: Hydrograph separation for storm event on 9/16/08.

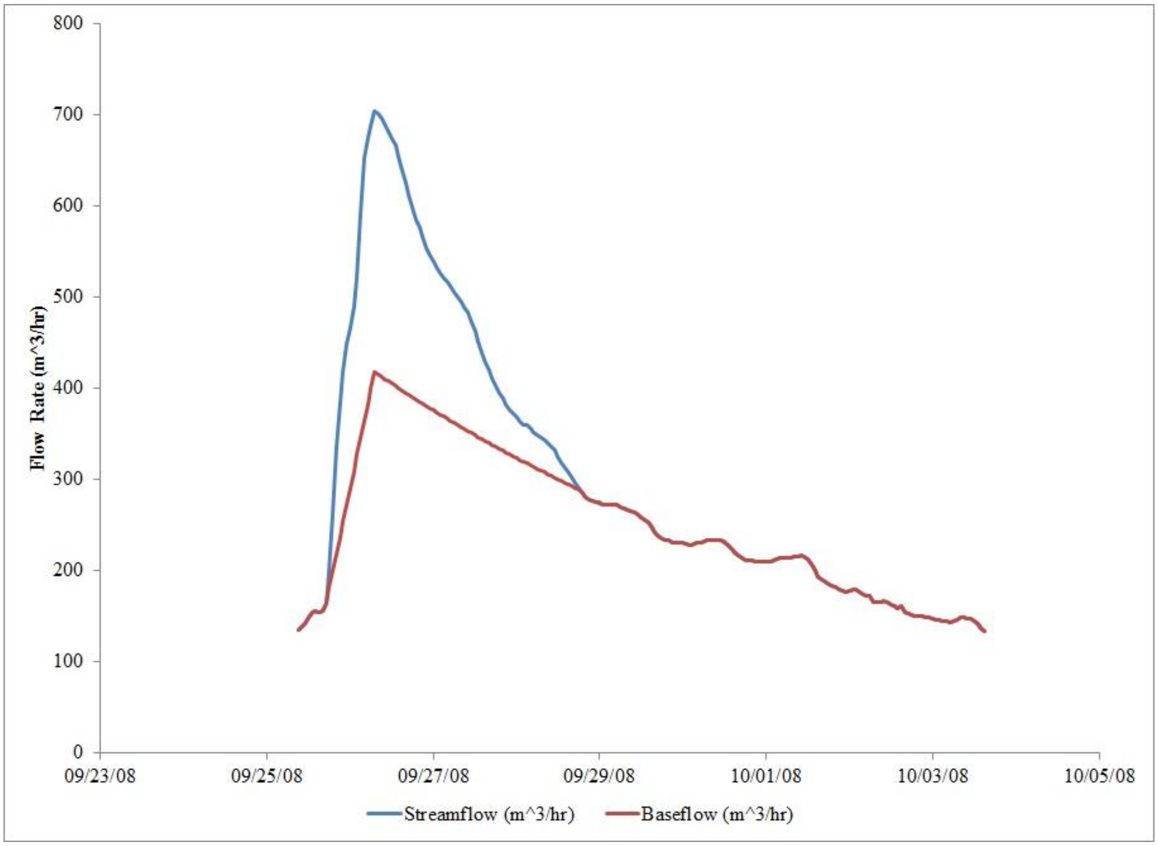


Figure A-5: Hydrograph separation for storm event on 9/25/08.

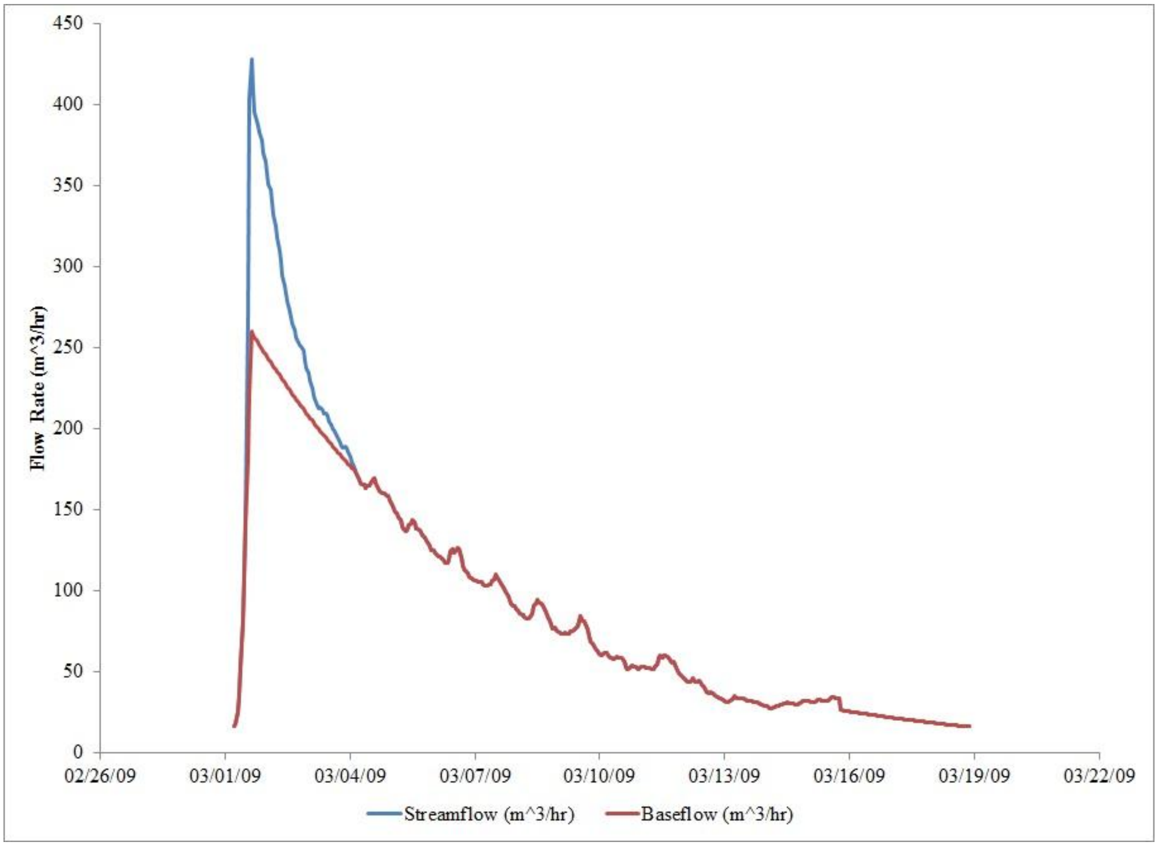
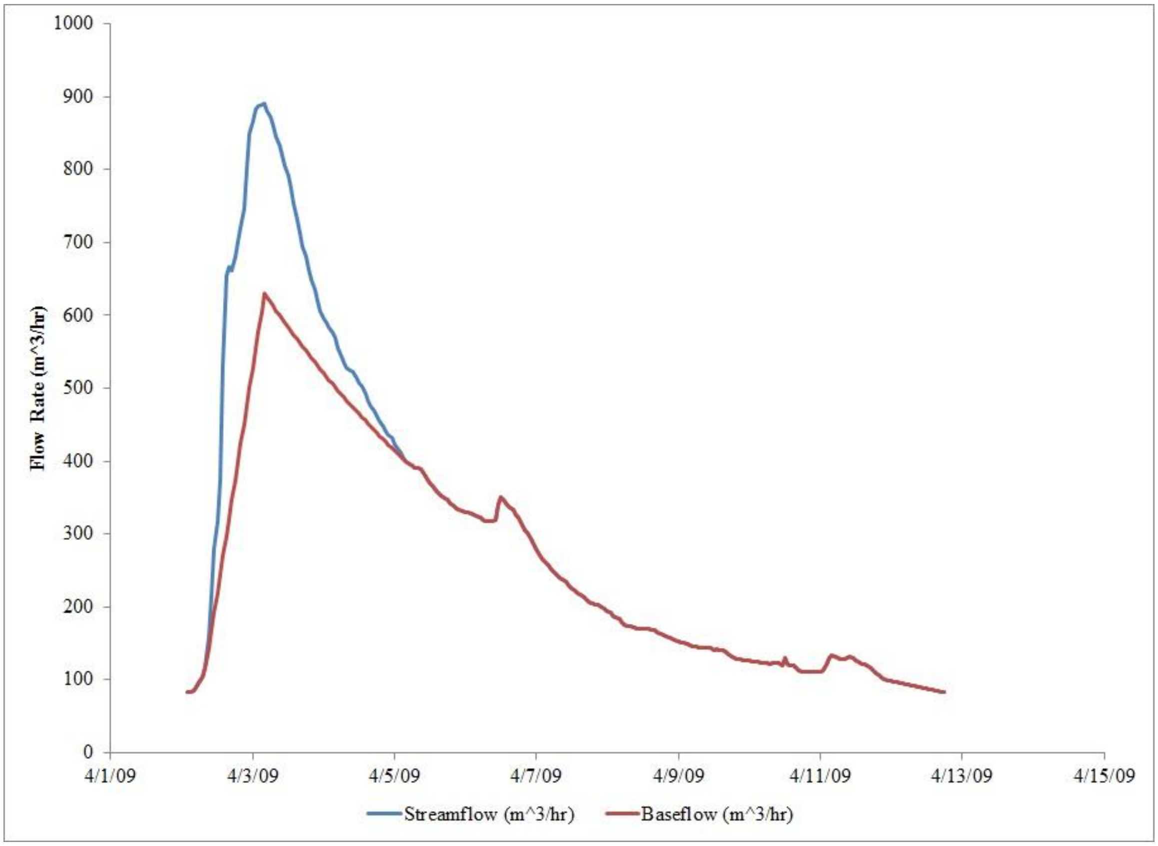


Figure A-6: Hydrograph separation for storm event on 3/1/09.



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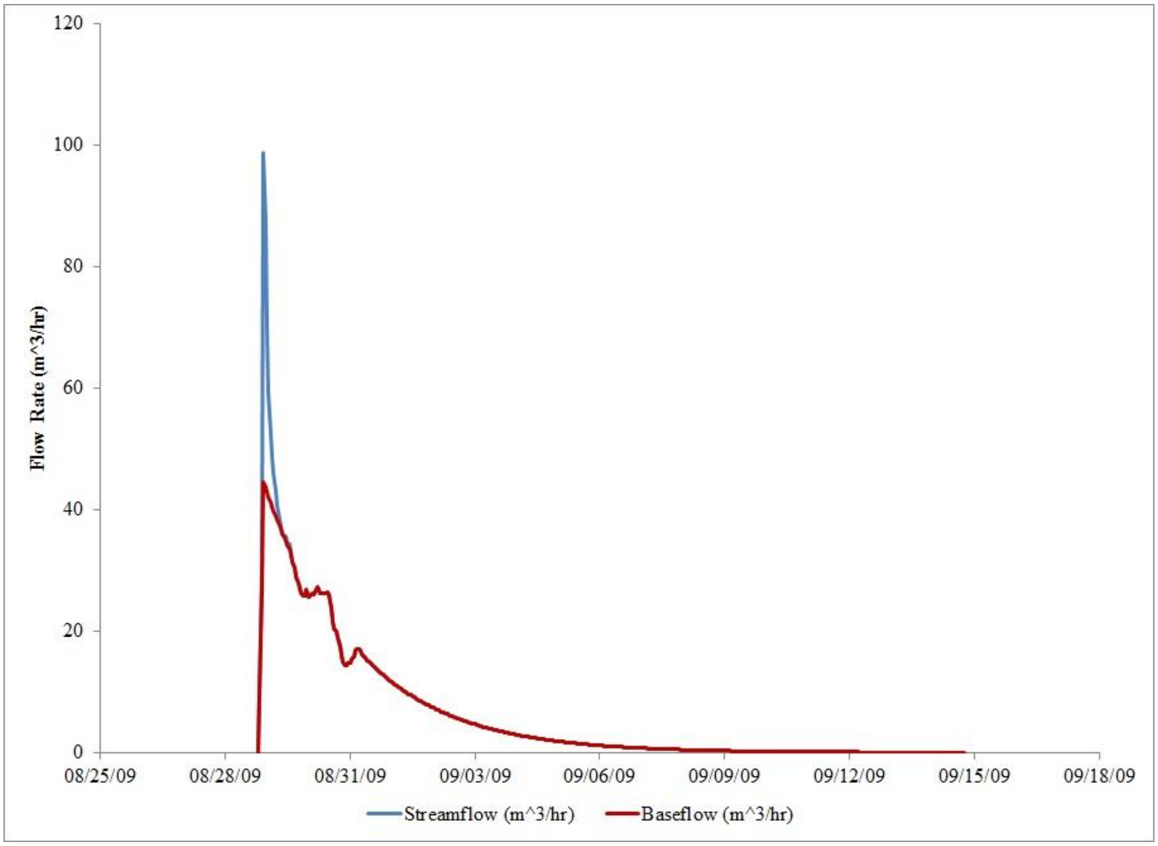


Figure A-8: Hydrograph separation for storm event on 8/28/09.

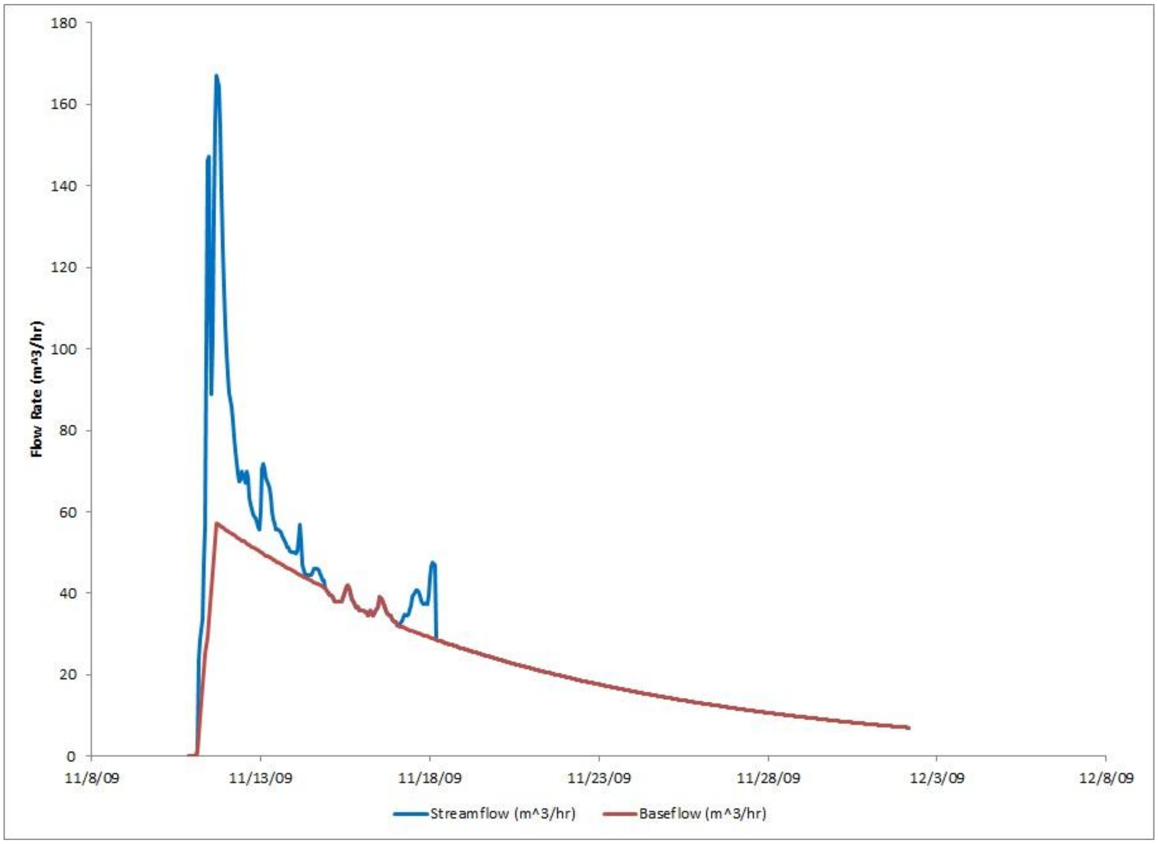


Figure A-9: Hydrograph separation for storm event on 11/10/09.

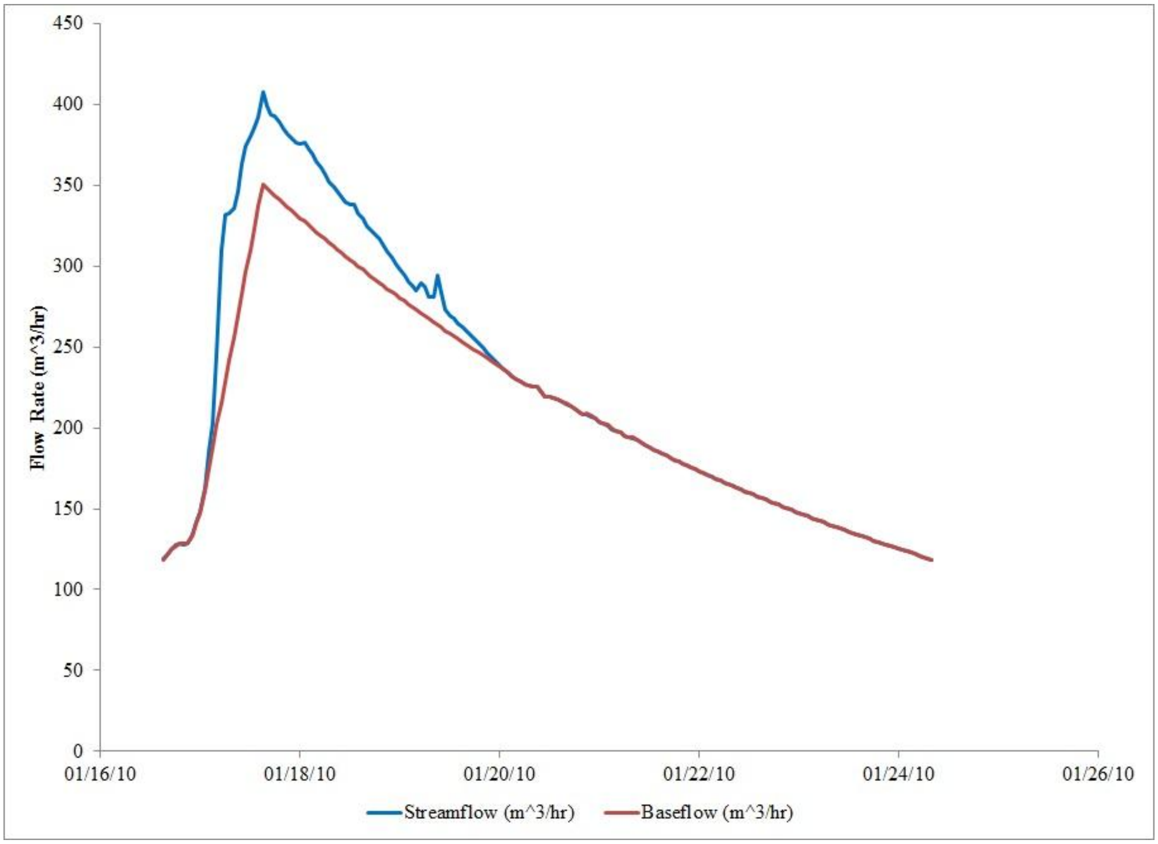


Figure A-10: Hydrograph separation for storm event on 1/16/10.

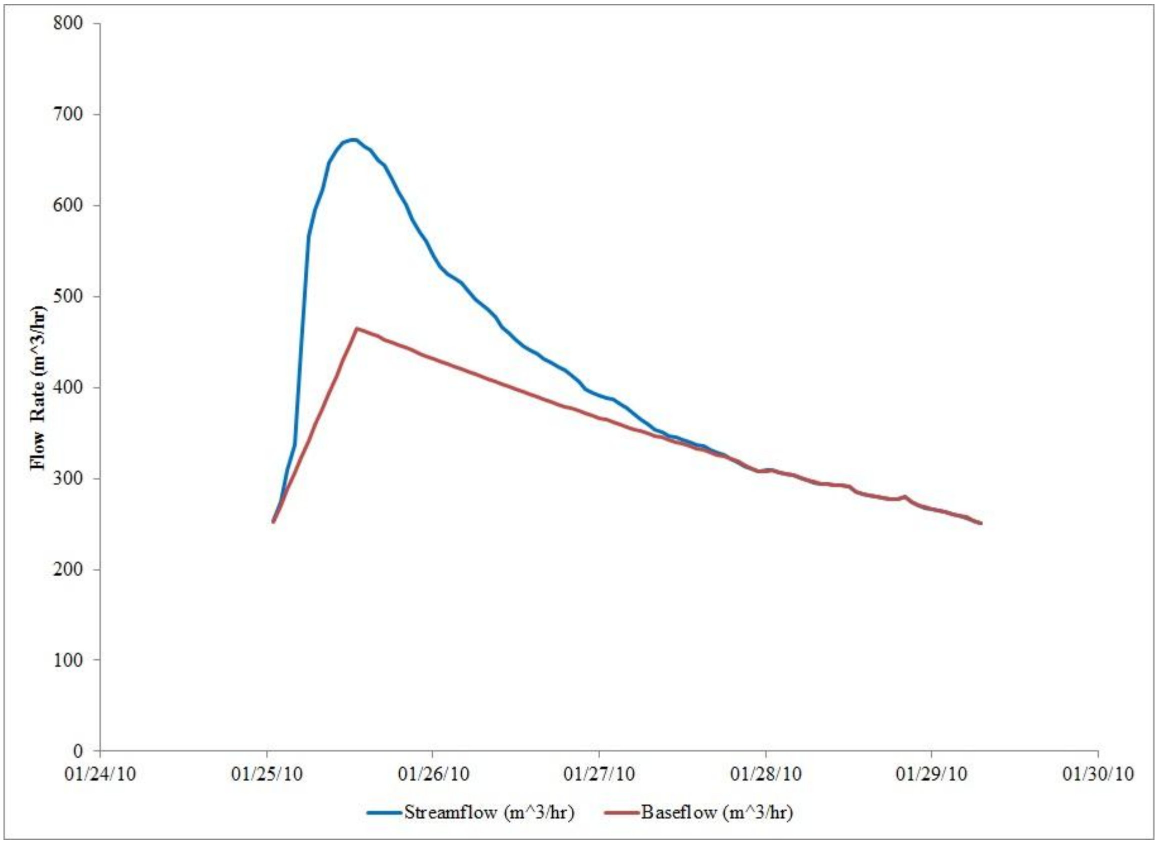


Figure A-11: Hydrograph separation for storm event on 1/25/10.

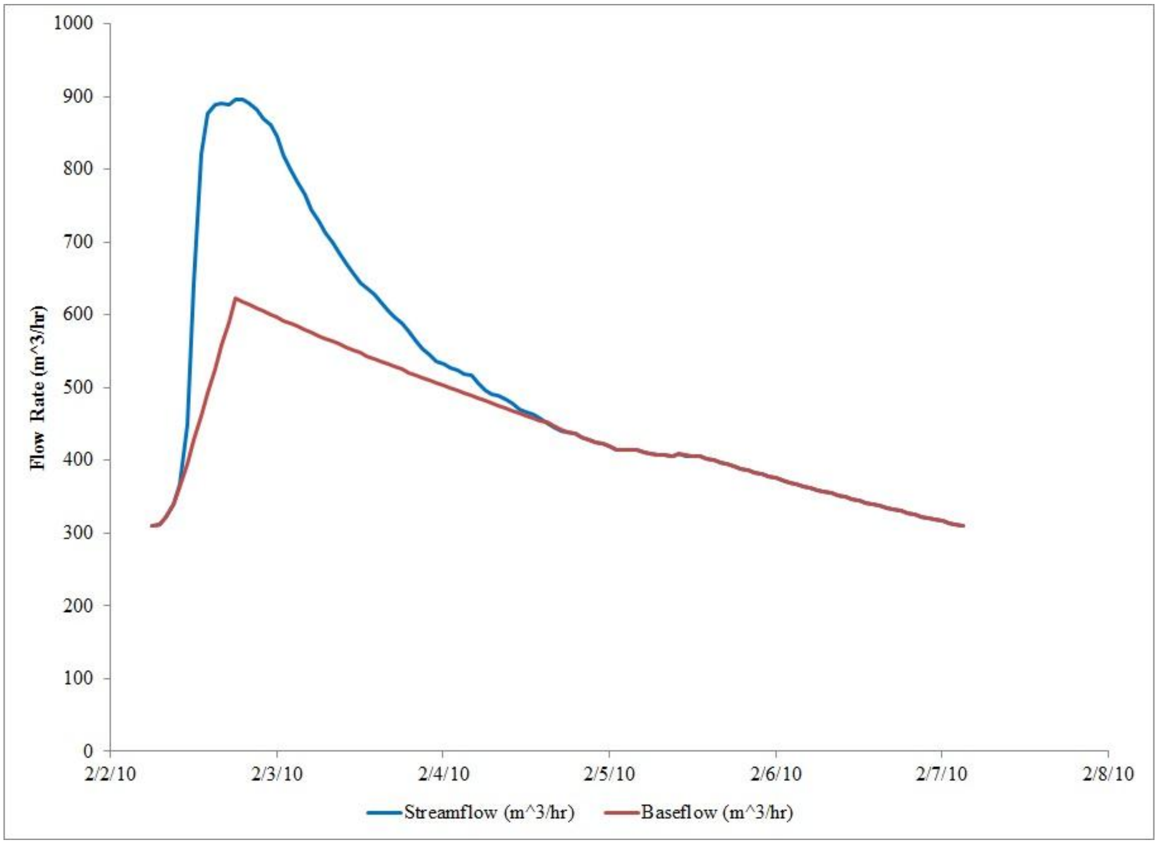


Figure A-12: Hydrograph separation for storm event on 2/2/10.

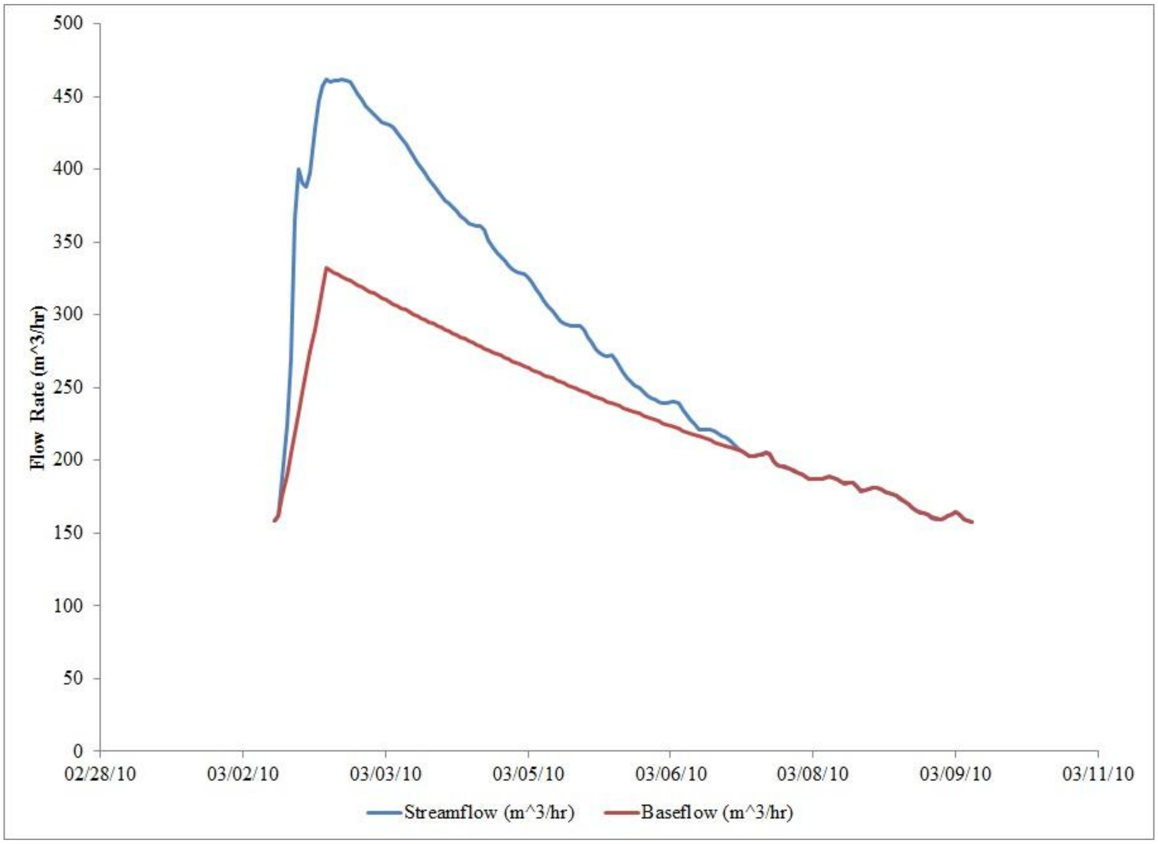


Figure A-13: Hydrograph separation for storm event on 3/2/10.

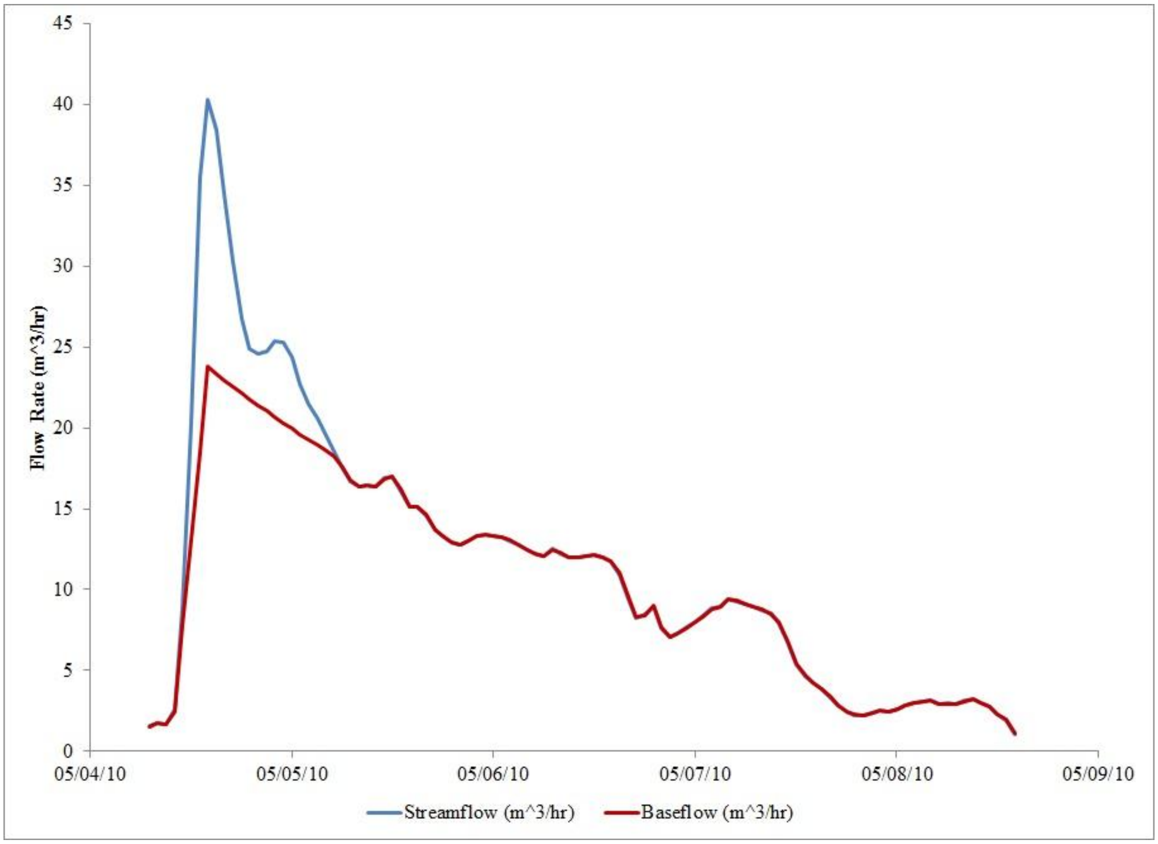


Figure A-14: Hydrograph separation for storm event on 5/4/10.

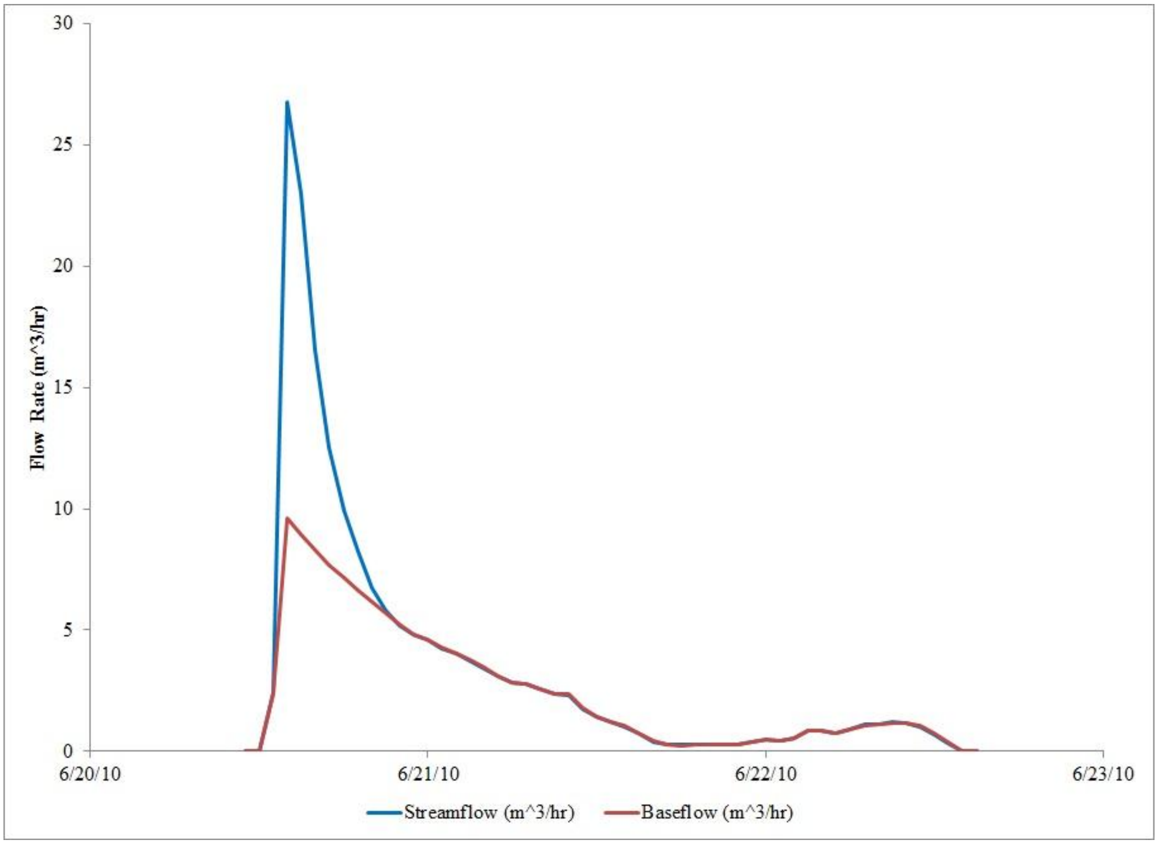
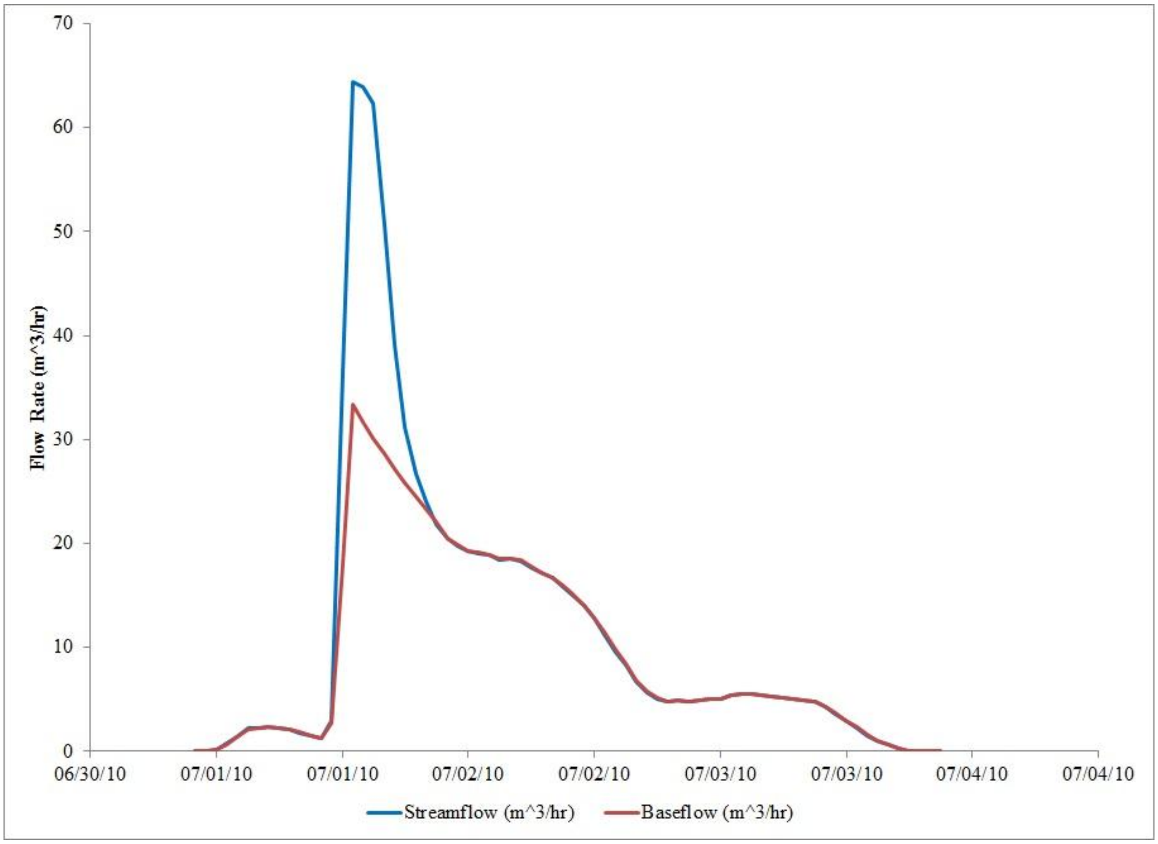


Figure A-15: Hydrograph separation for storm event on 6/20/10.



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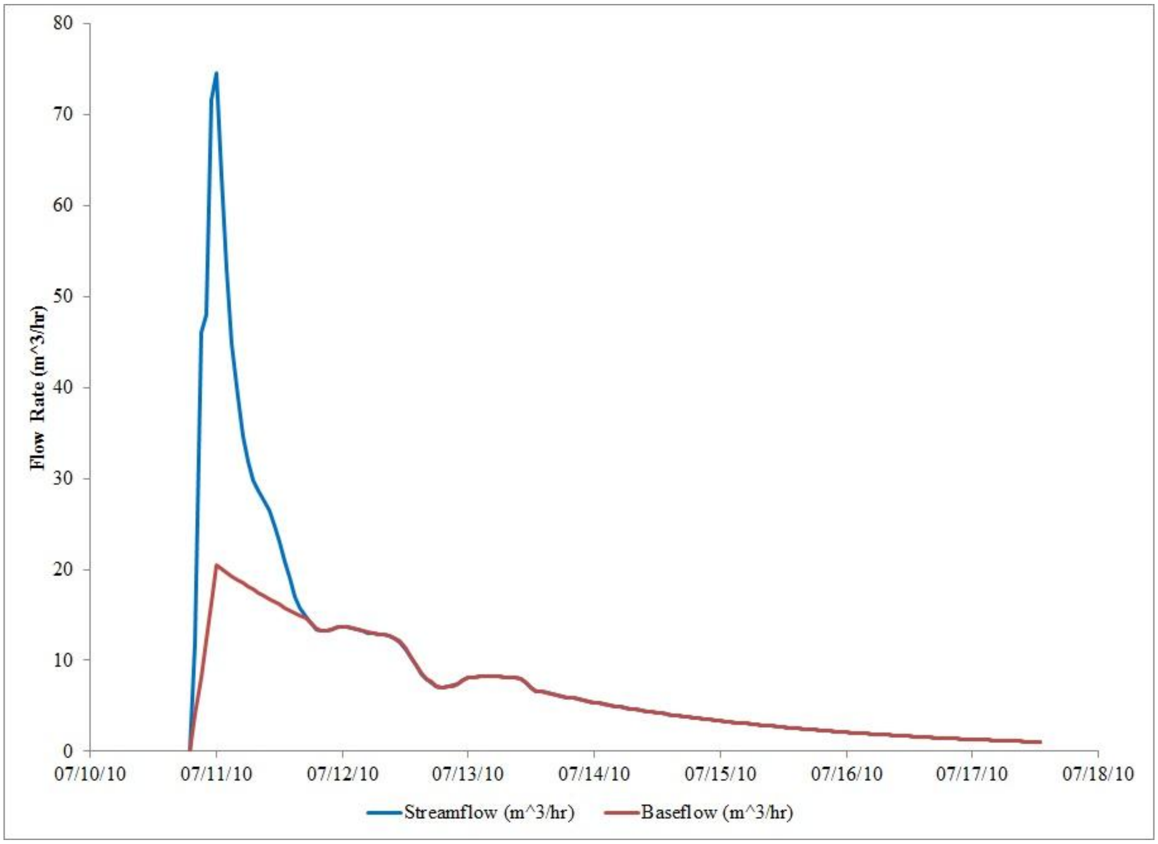


Figure A-17: Hydrograph separation for storm event on 7/10/10.

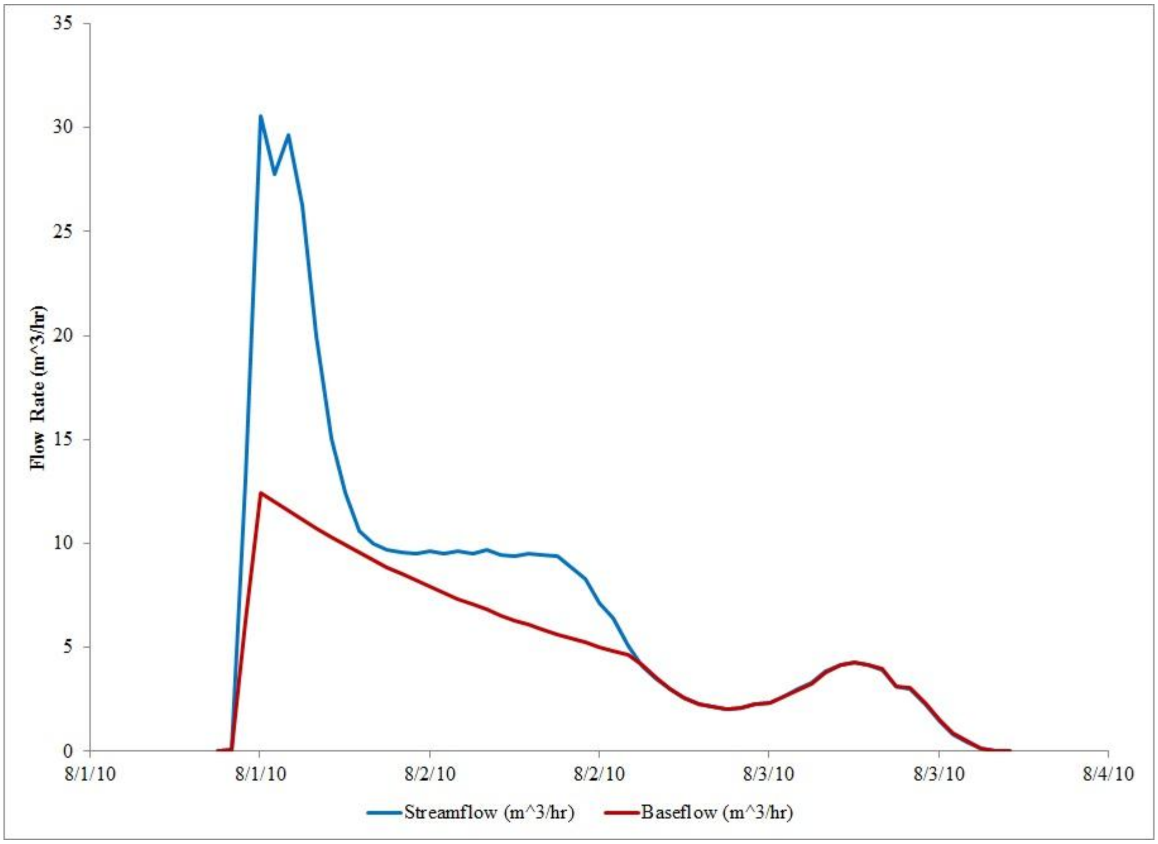


Figure A-18: Hydrograph separation for storm event on 8/1/10.

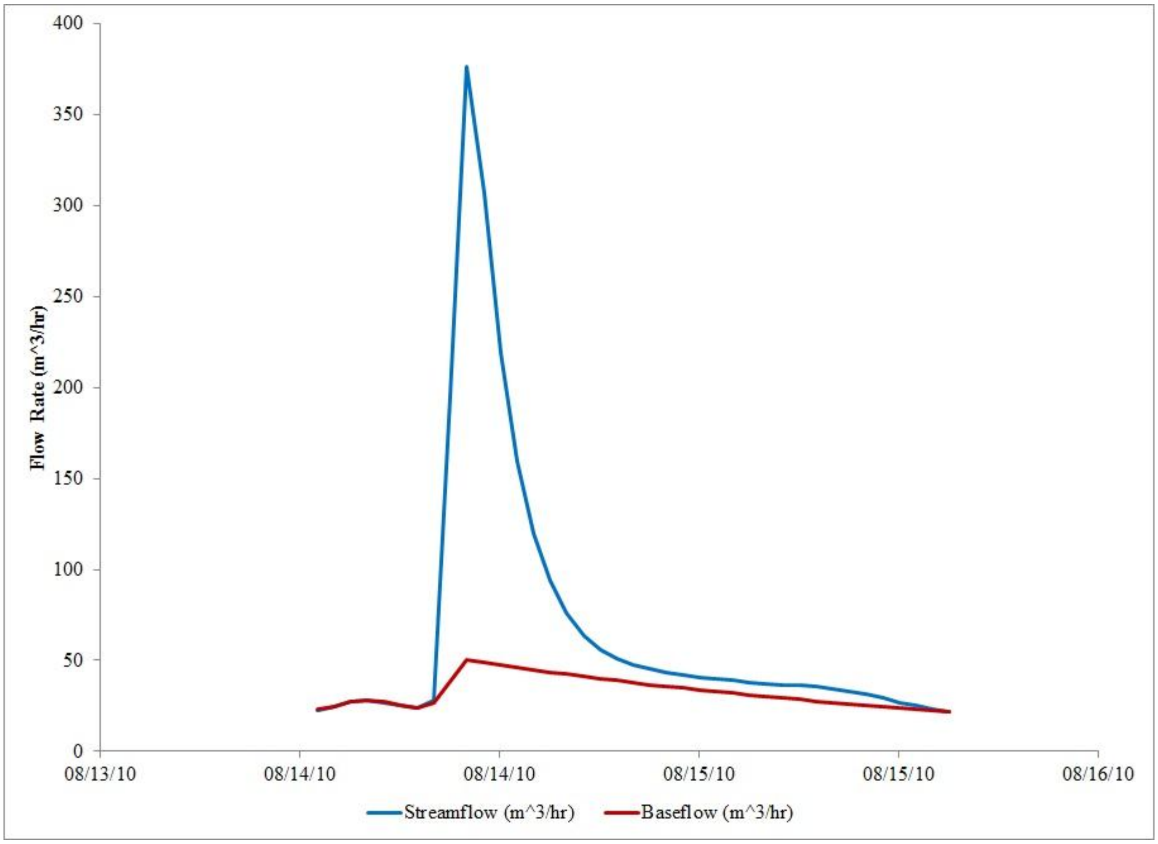


Figure A-19: Hydrograph separation for storm event on 8/13/10.

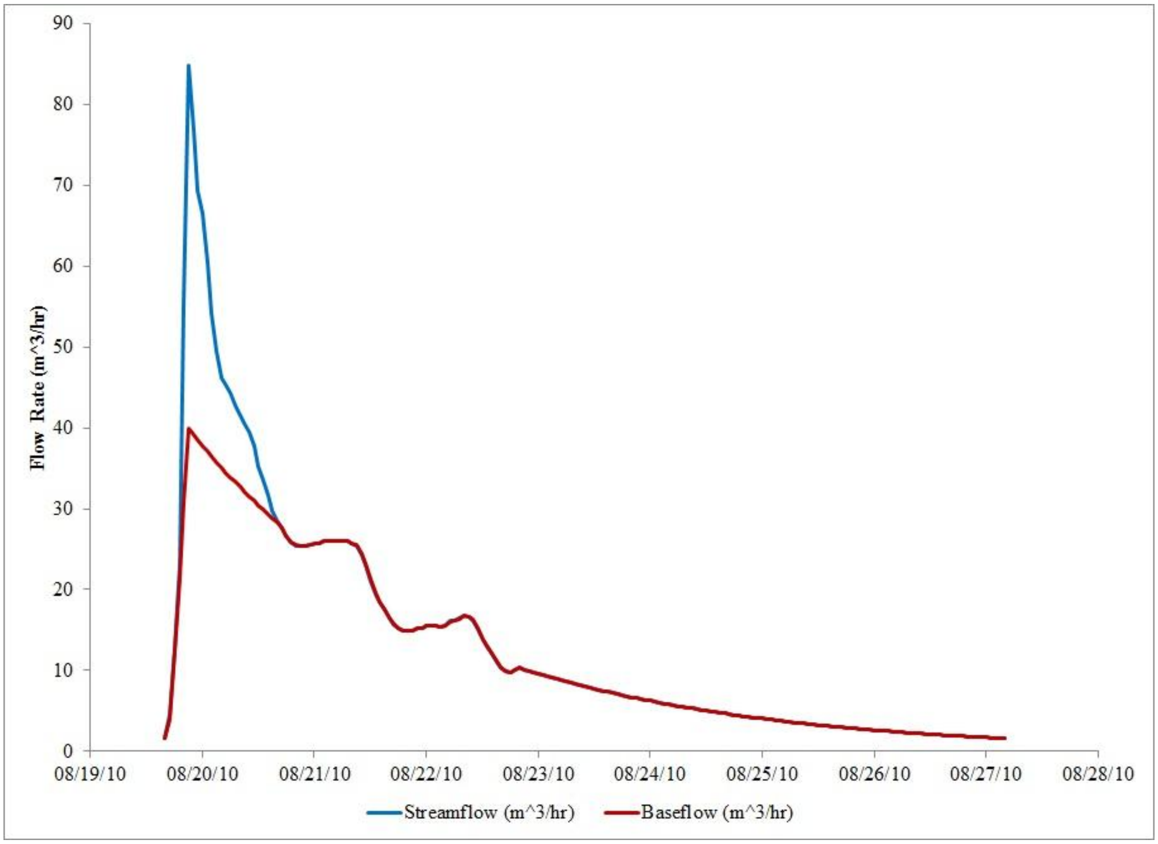


Figure A-20: Hydrograph separation for storm event on 8/19/10.

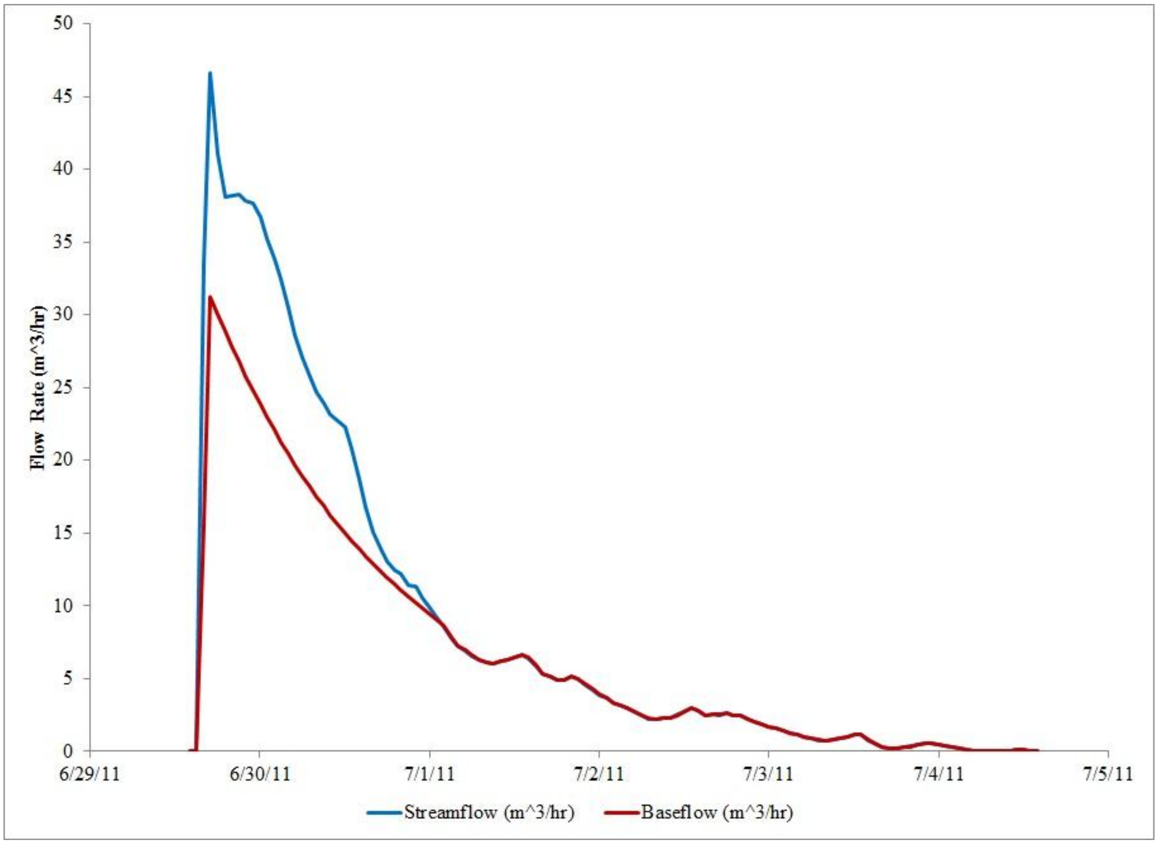


Figure A-21: Hydrograph separation for storm event on 6/29/11.

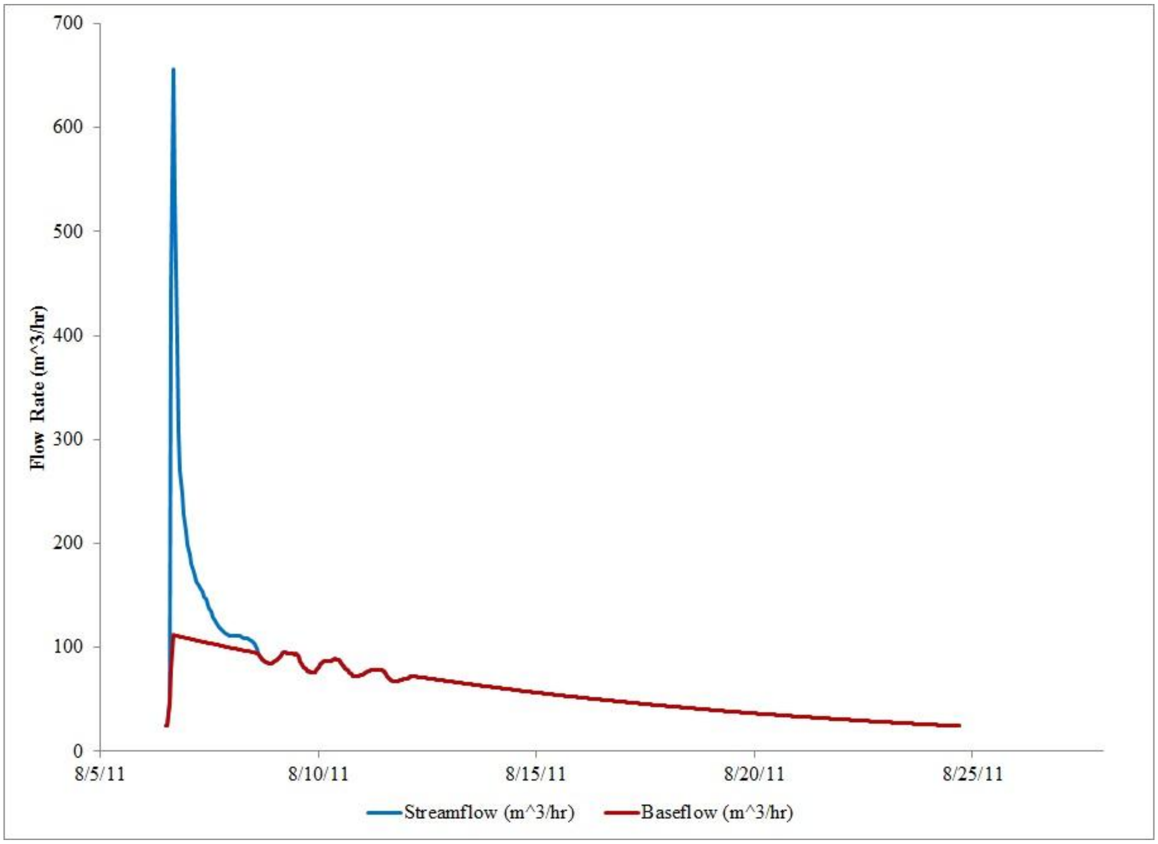


Figure A-22: Hydrograph separation for storm event on 8/6/11.

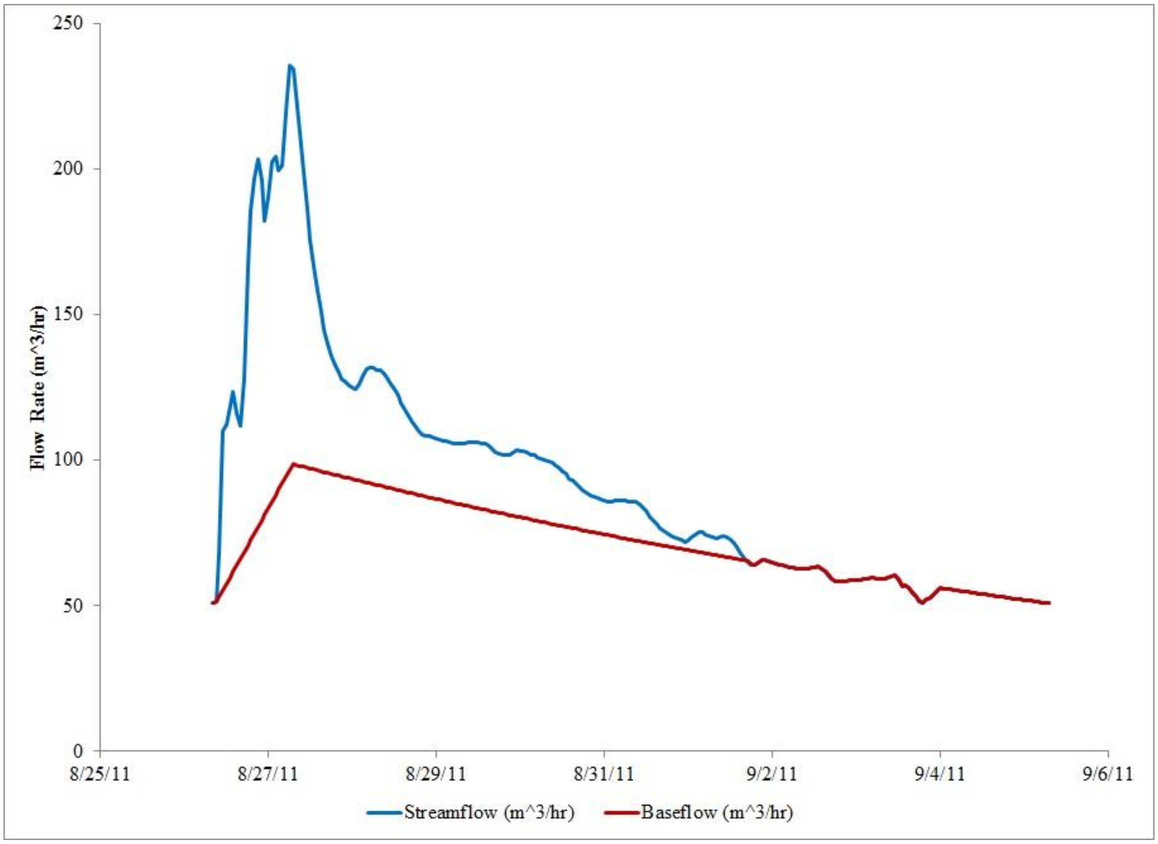


Figure A-23: Hydrograph separation for storm event on 8/25/11.

Appendix B

Storm Event Hydrograph Separations for Watershed 80

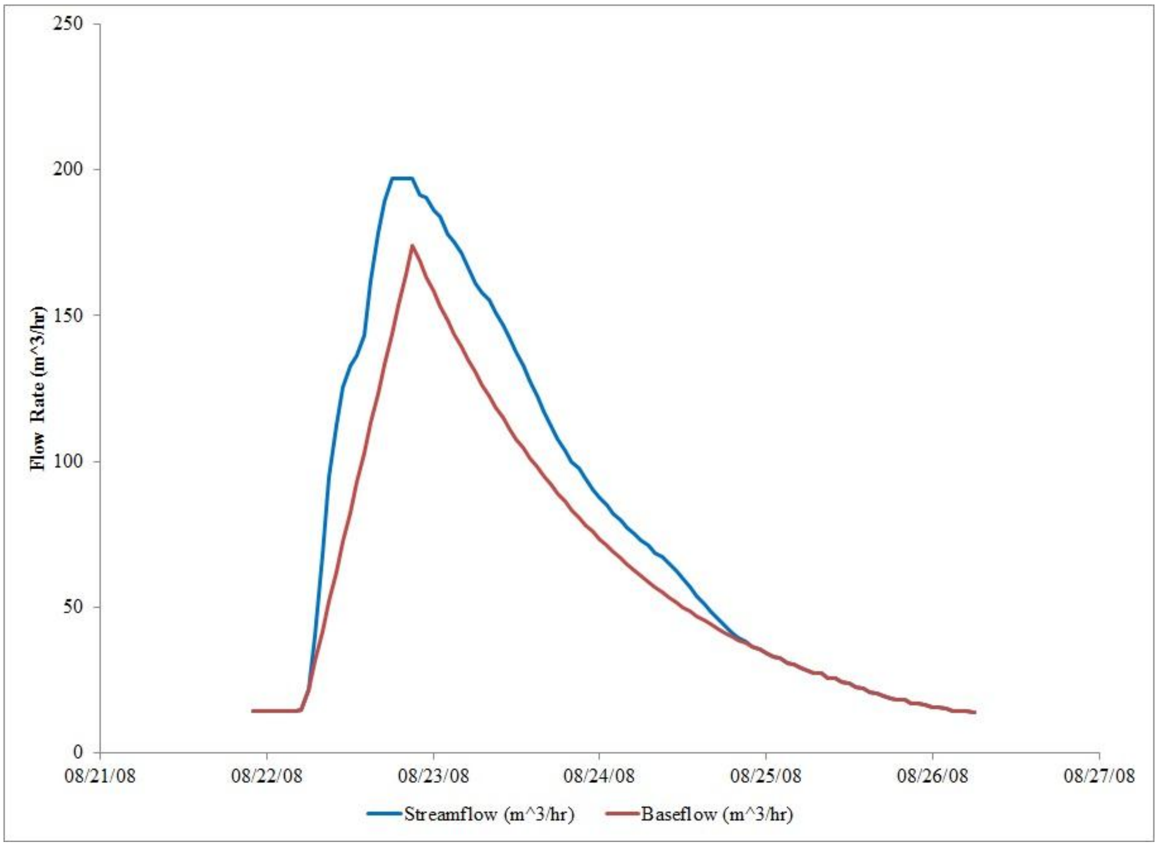


Figure B-1: Hydrograph separation for storm event on 8/21/08.

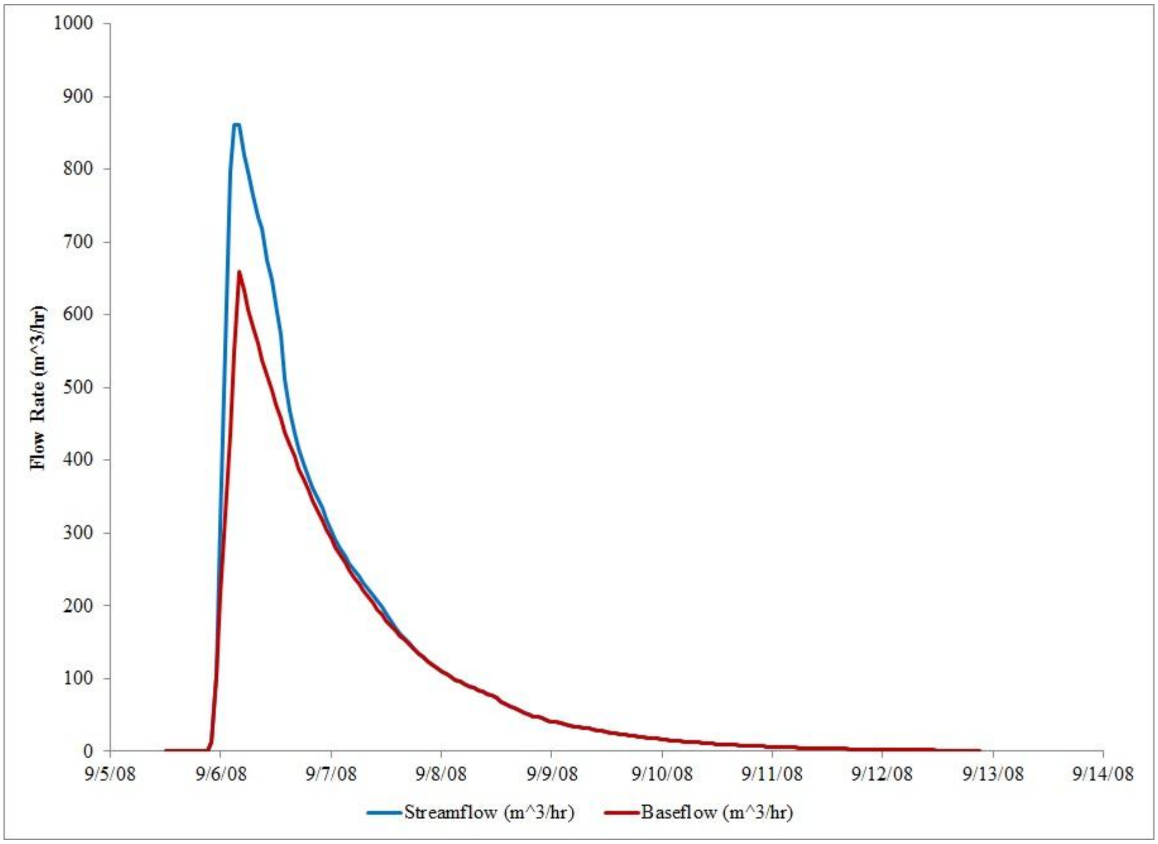
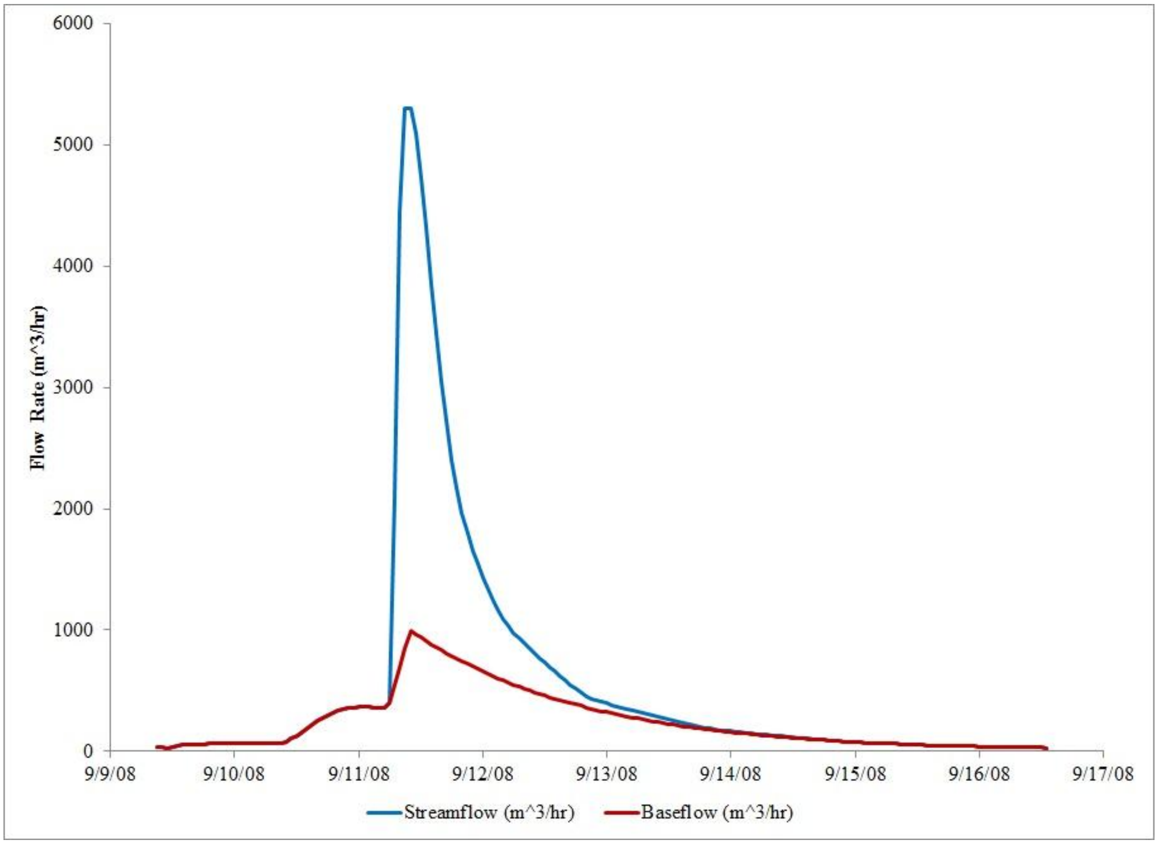


Figure B-2: Hydrograph separation for storm event on 9/5/08.



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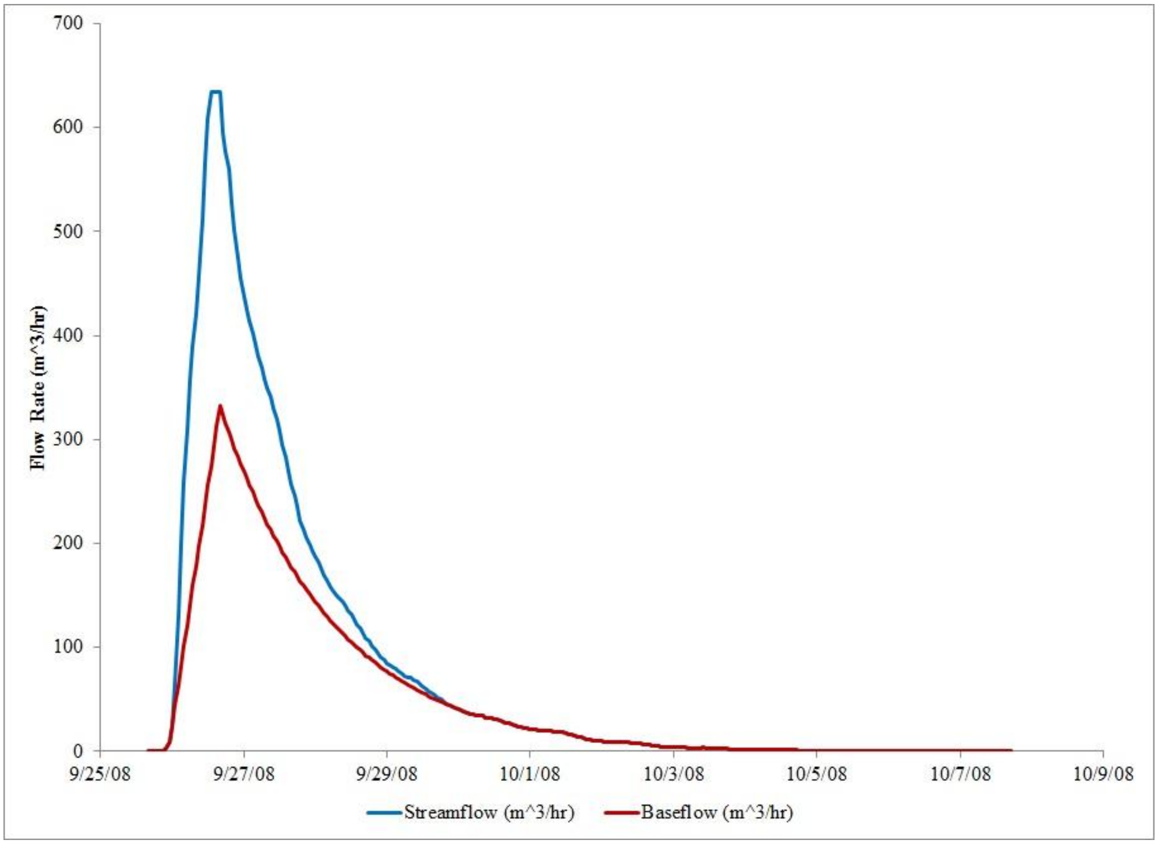


Figure B-4: Hydrograph separation for storm event on 9/25/08.

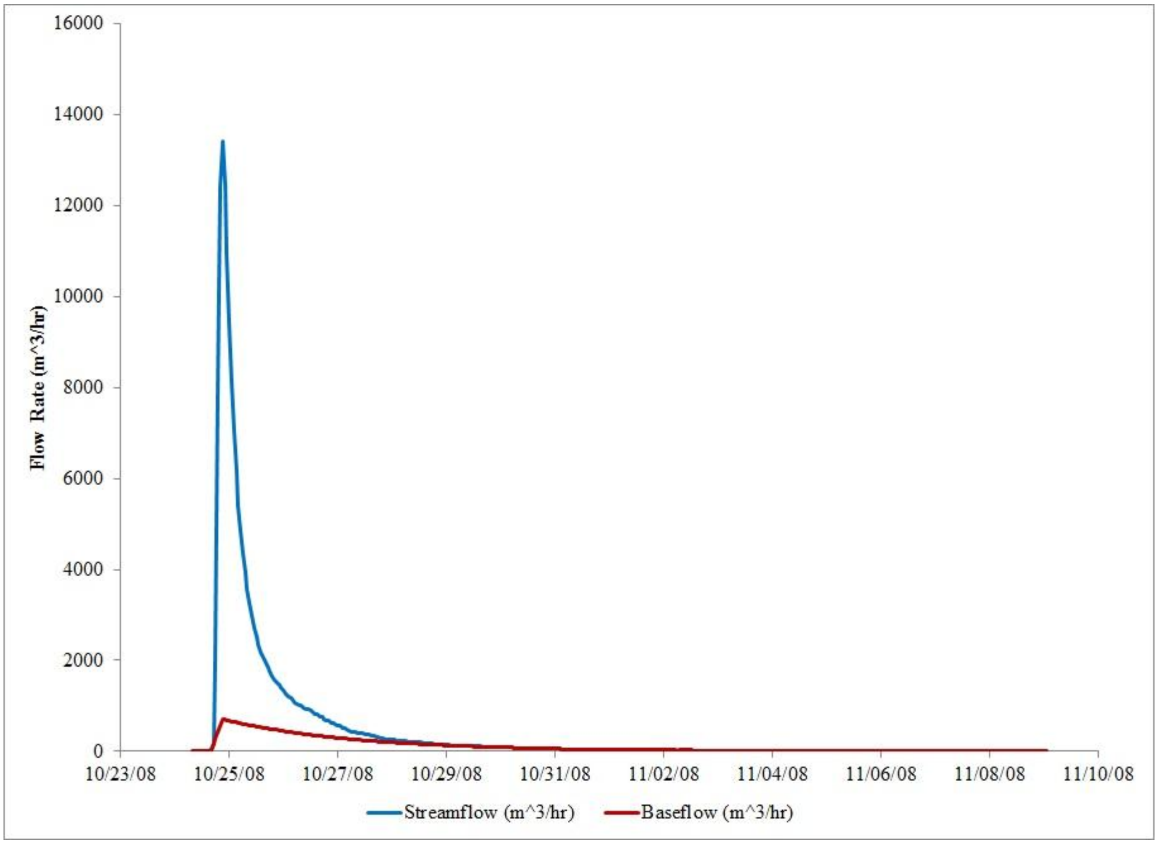


Figure B-5: Hydrograph separation for storm event on 10/24/08.

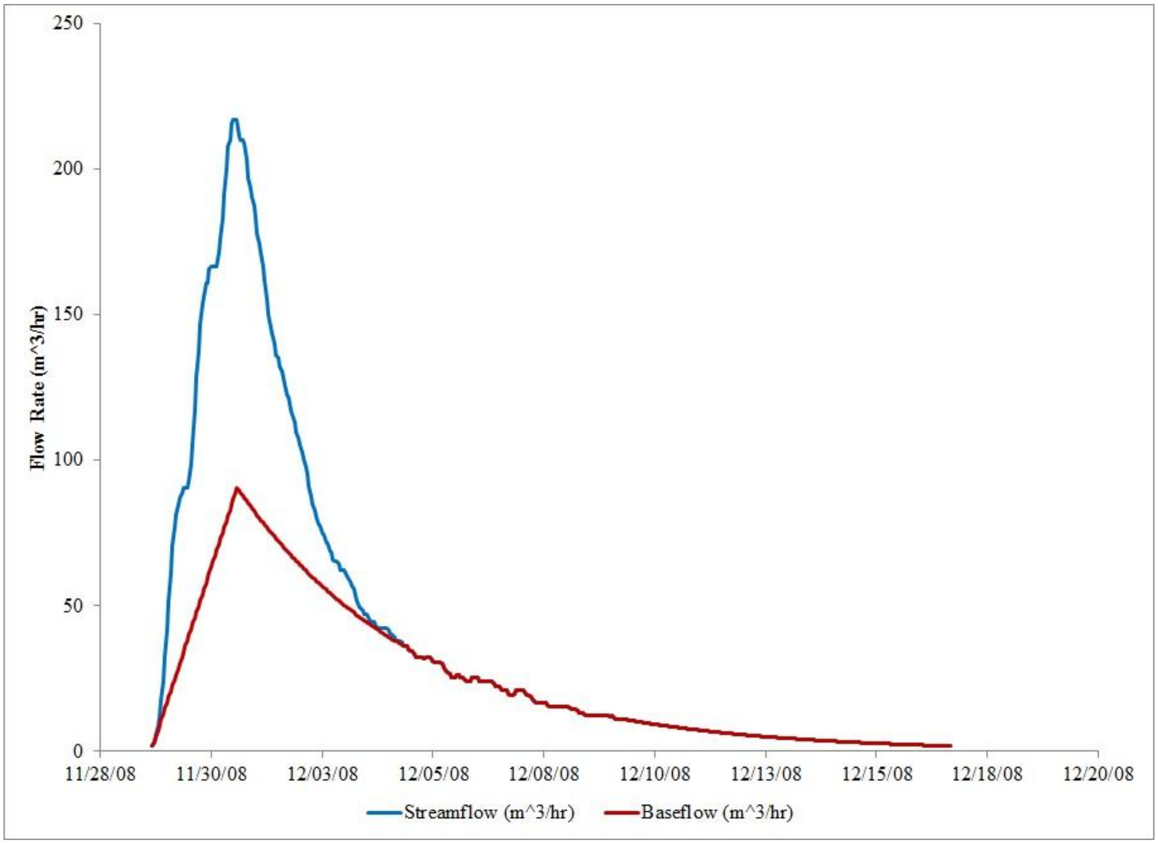
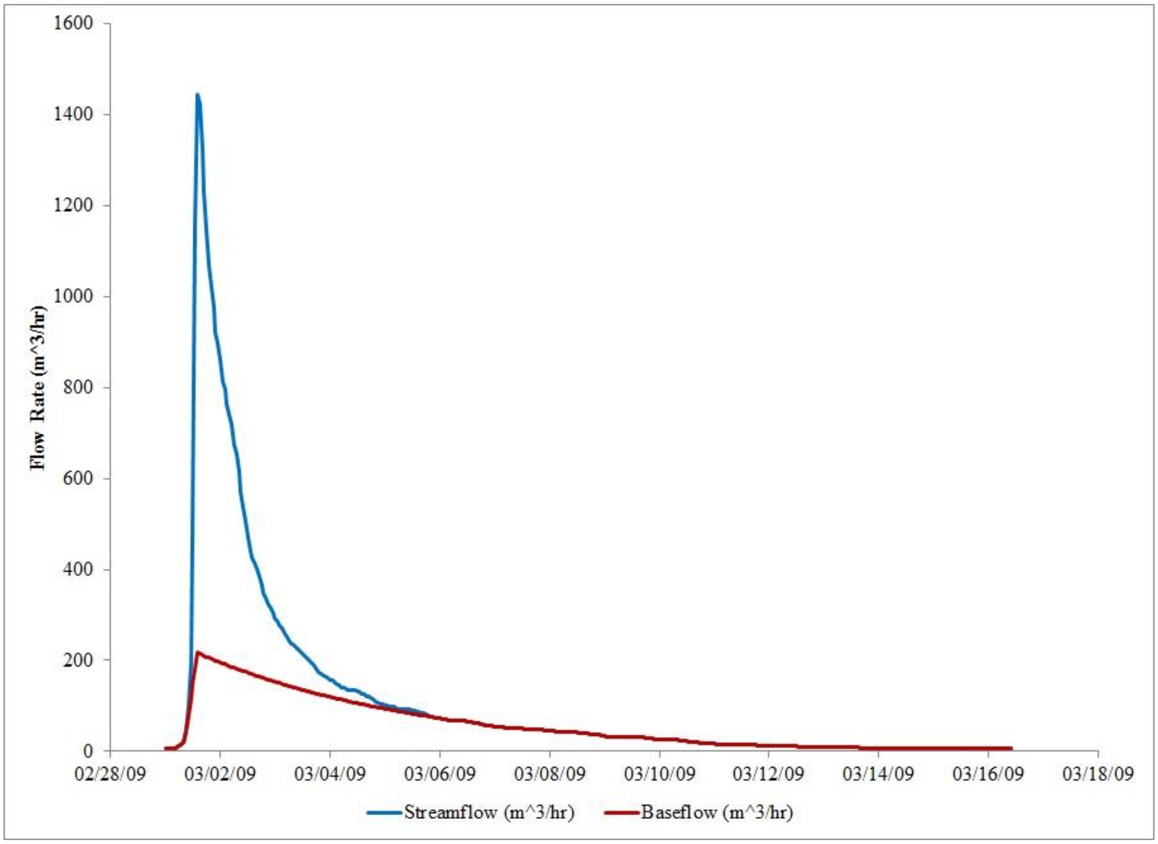


Figure B-6: Hydrograph separation for storm event on 11/29/08.



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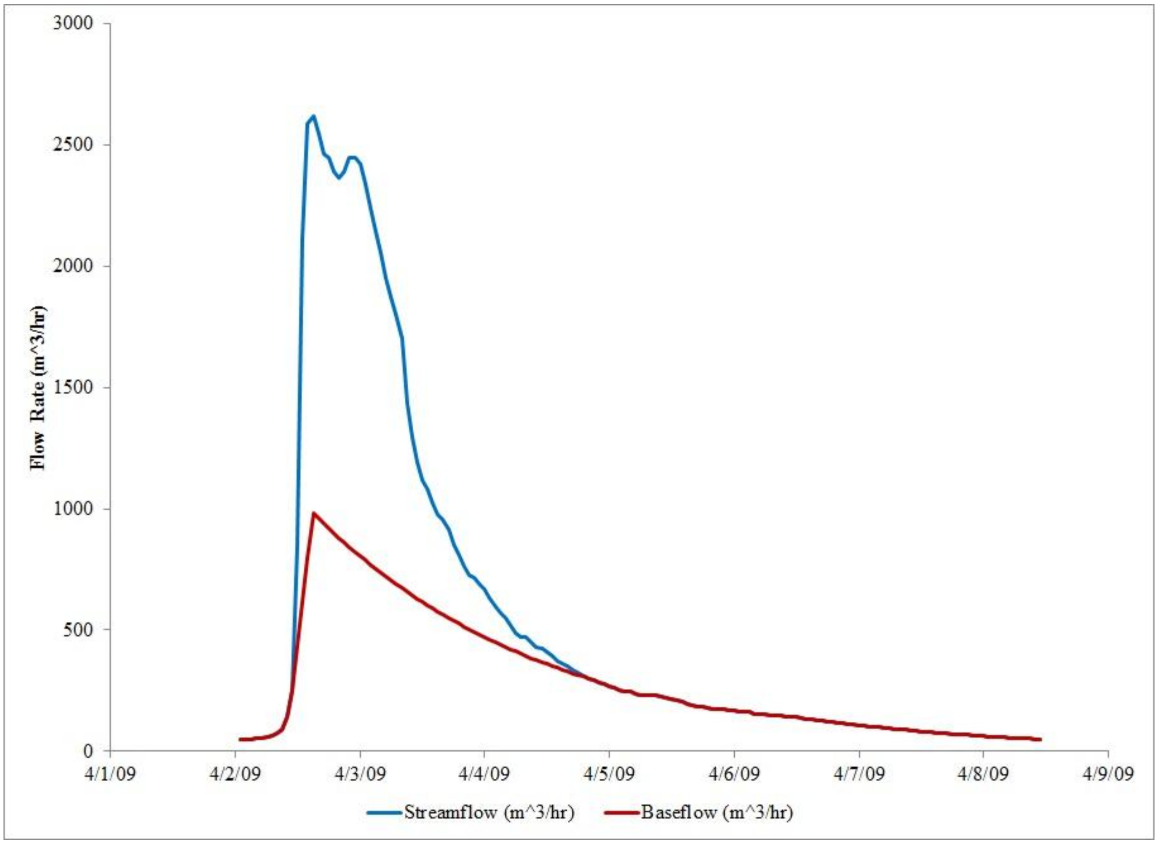

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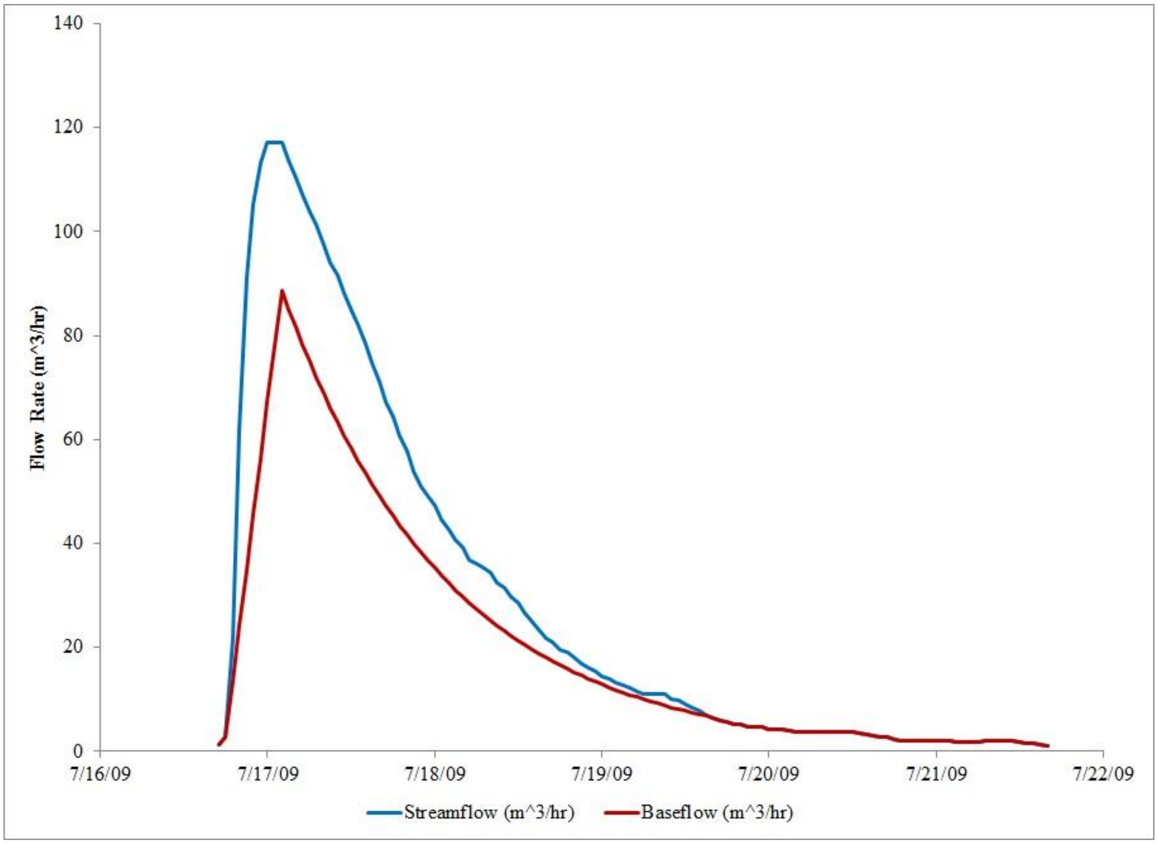
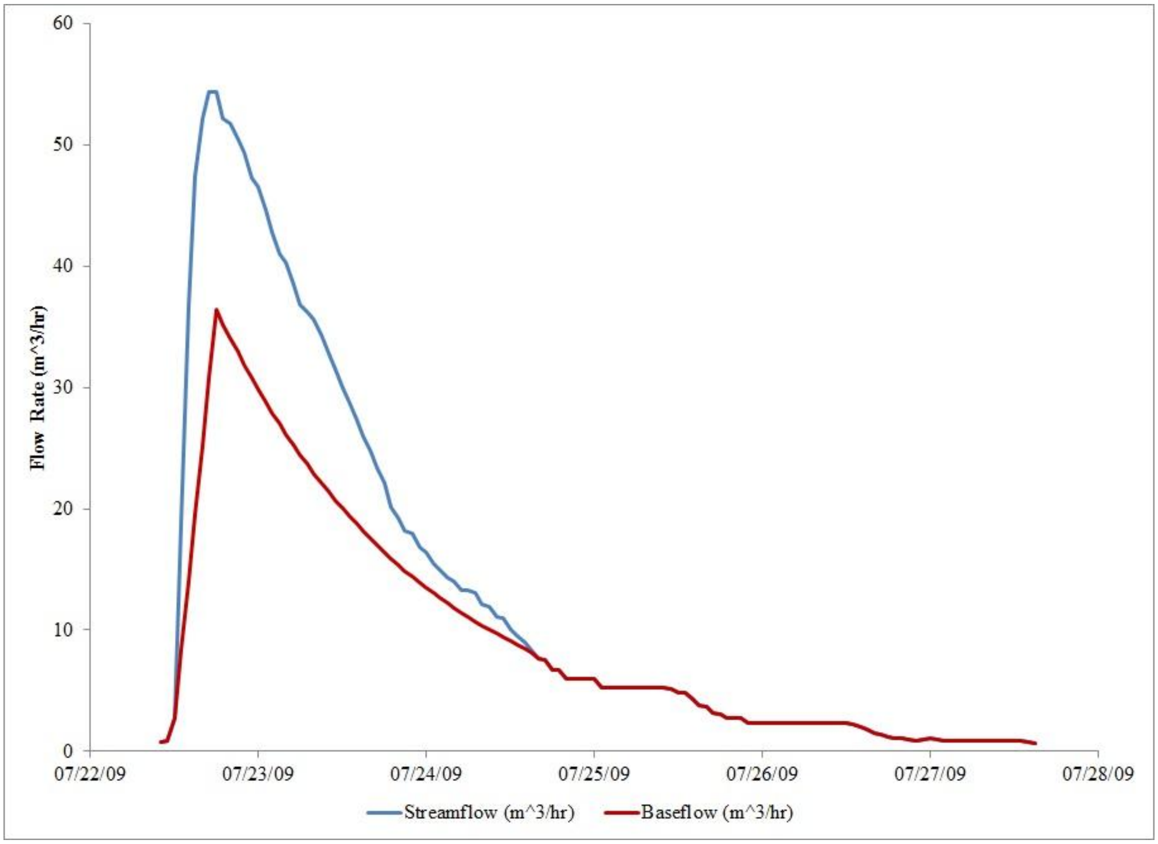


Figure B-9: Hydrograph separation for storm event on 7/16/09.



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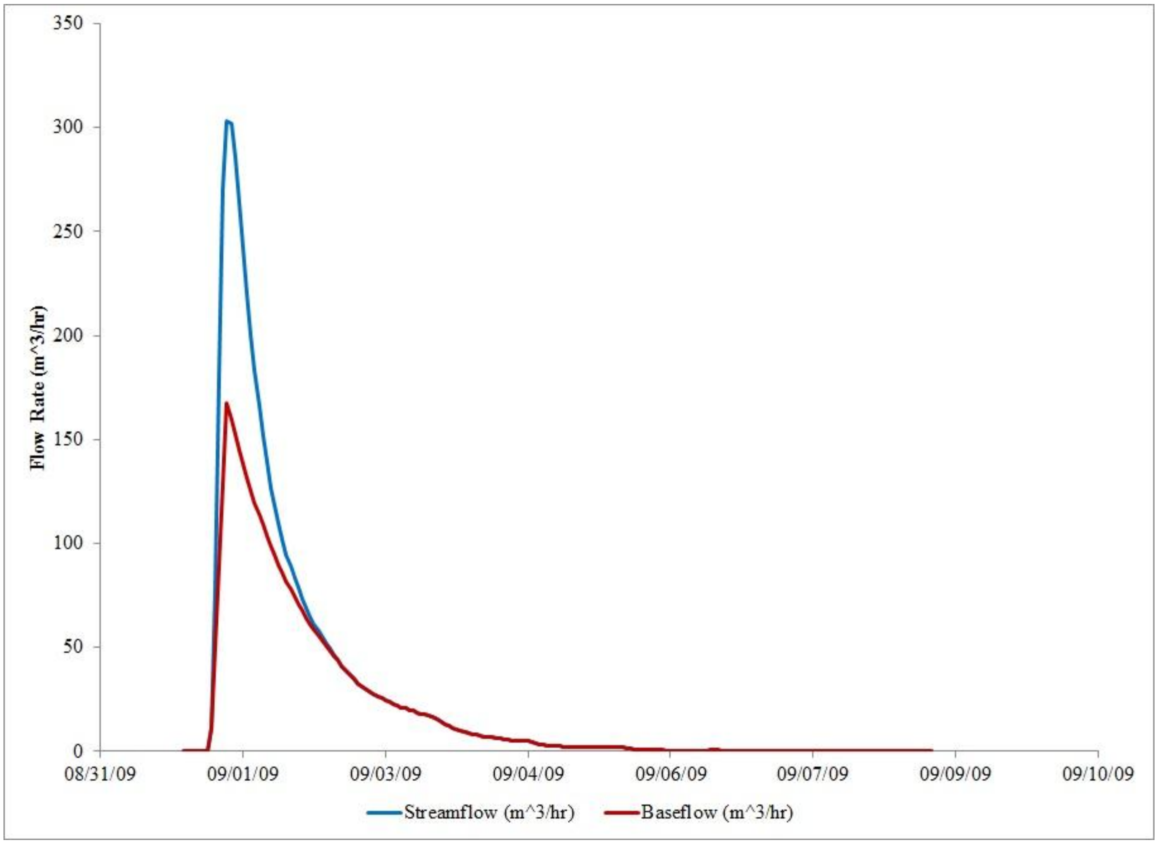

Figure B-11: Hydrograph separation for storm event on 8/31/09.

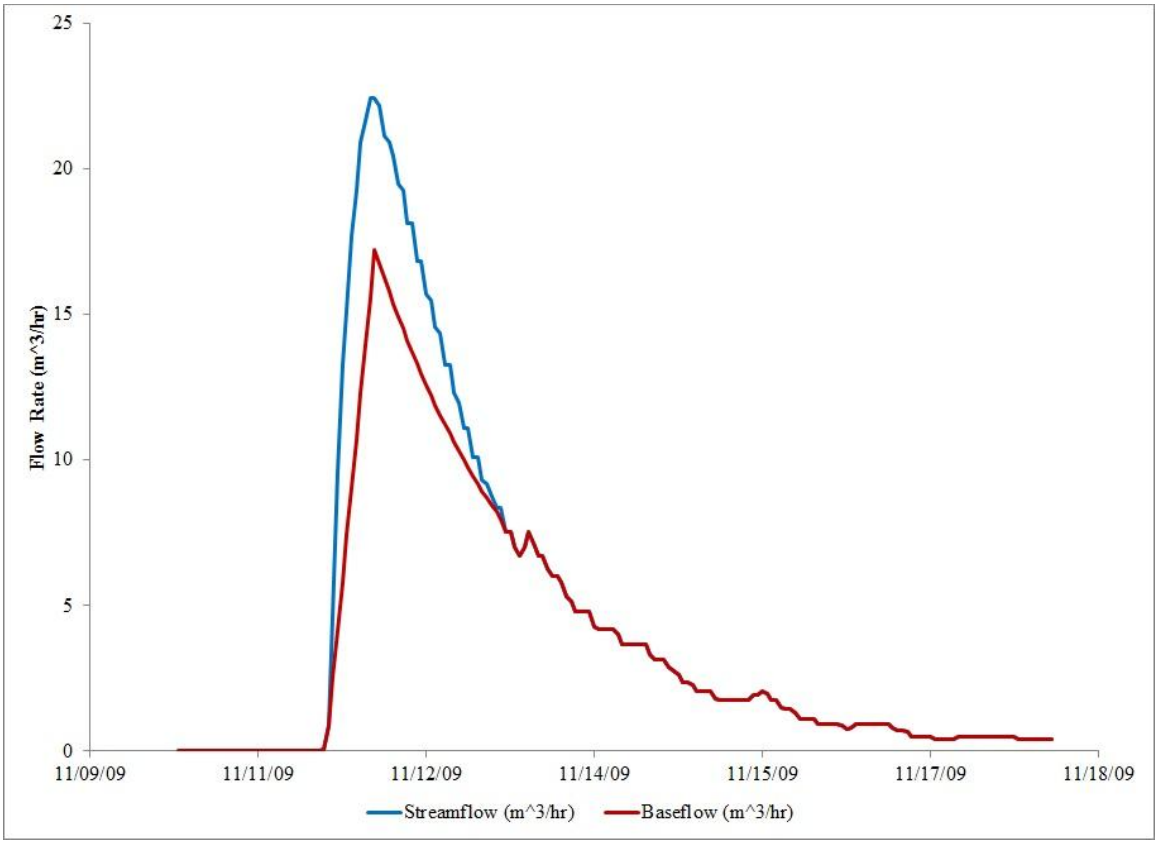

Figure B-12: Hydrograph separation for storm event on 11/11/09.

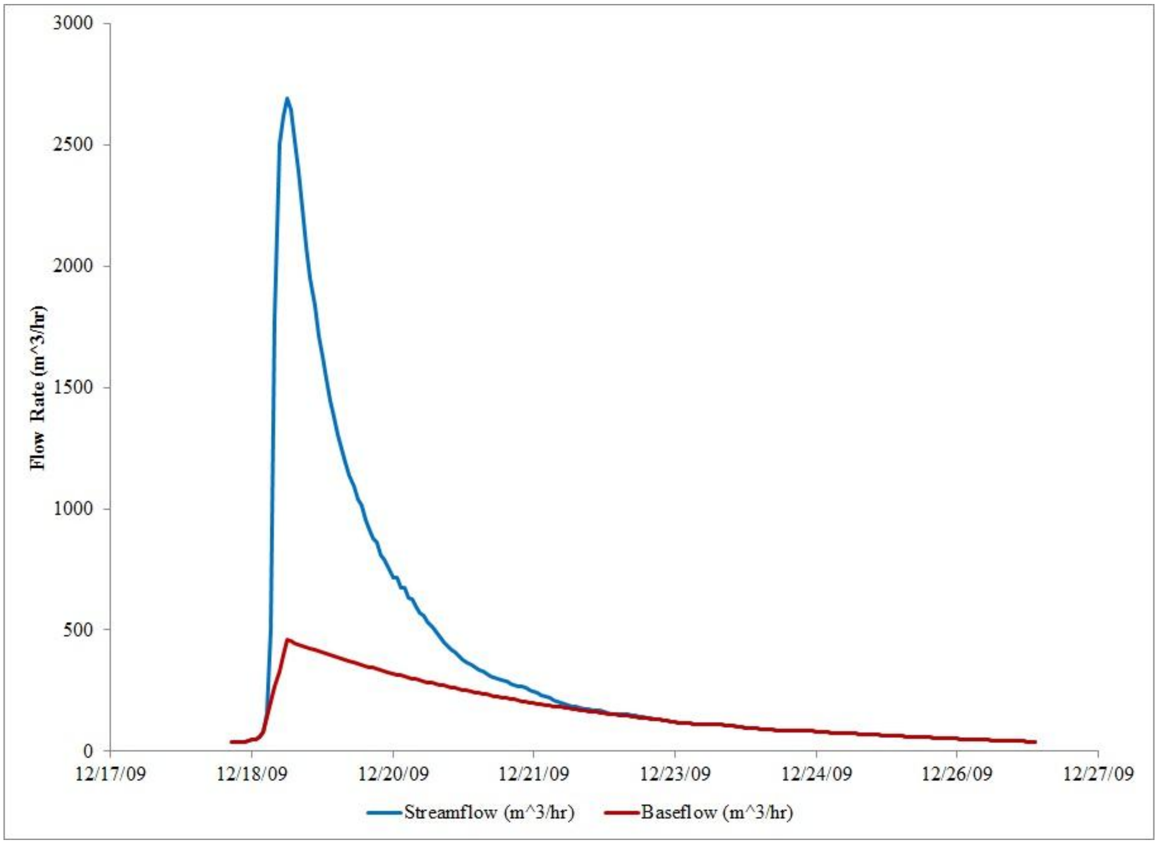

Figure B-13: Hydrograph separation for storm event on 12/18/09.

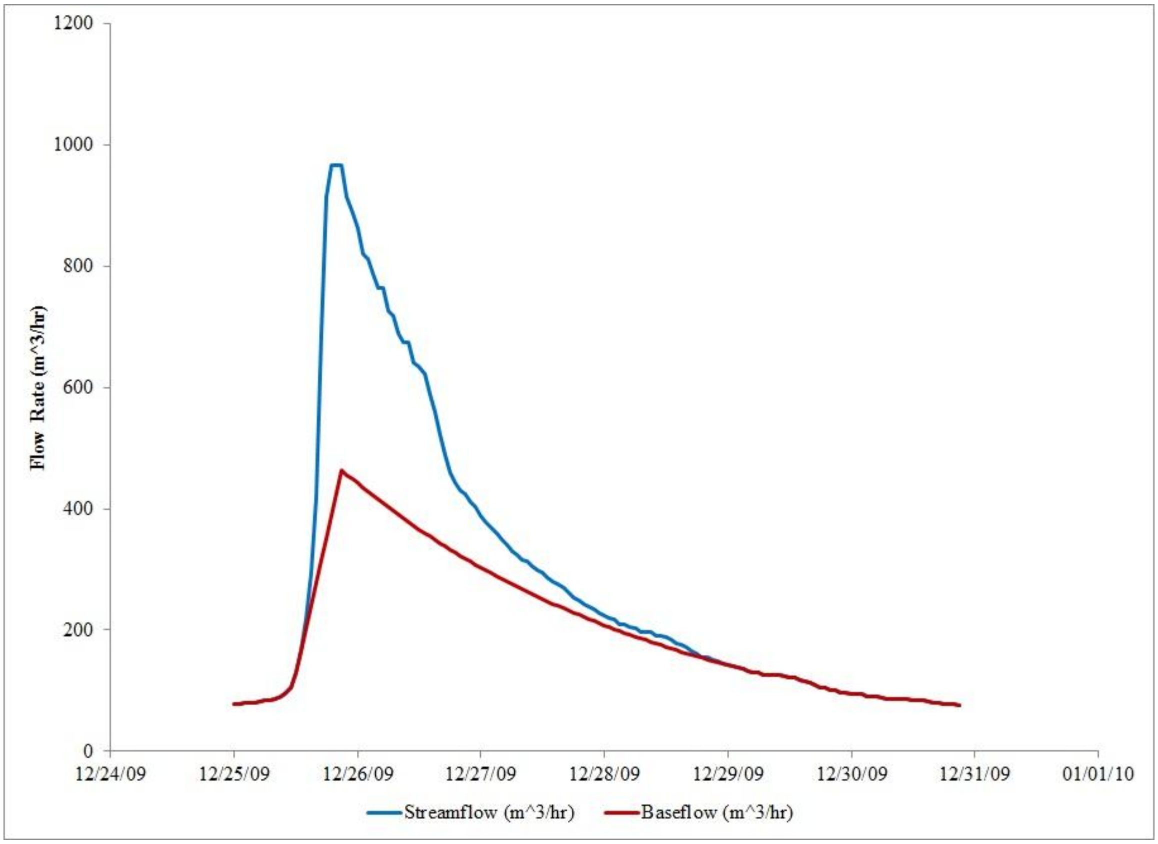

Figure B-14: Hydrograph separation for storm event on 12/25/09.

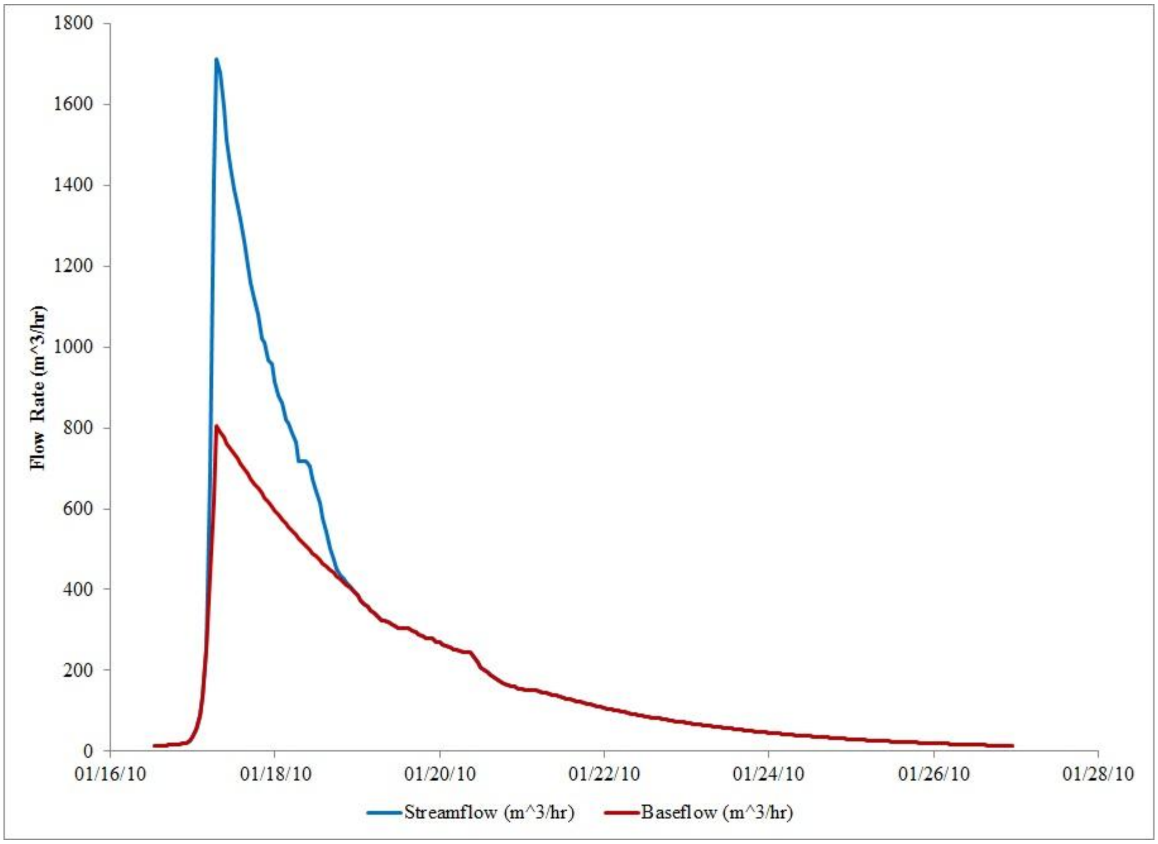

Figure B-15: Hydrograph separation for storm event on 1/16/10.

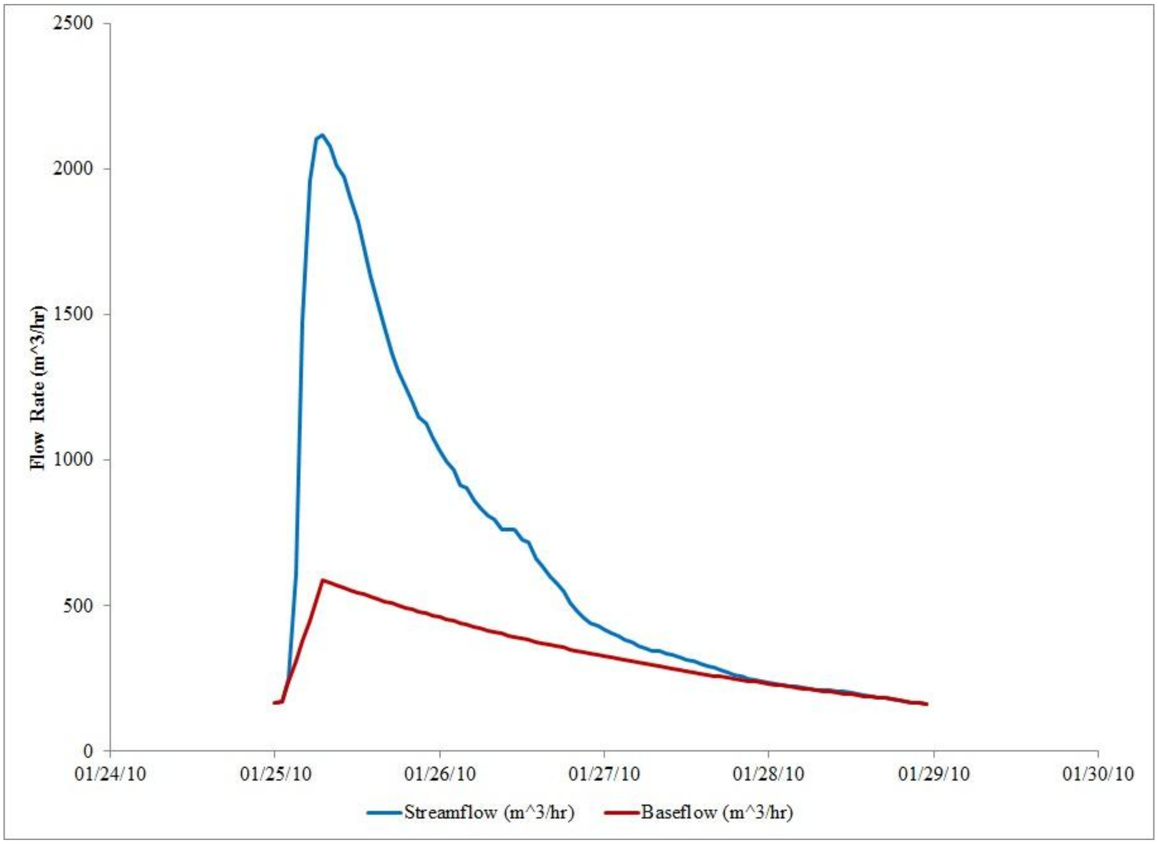

Figure B-16: Hydrograph separation for storm event on 1/25/10.

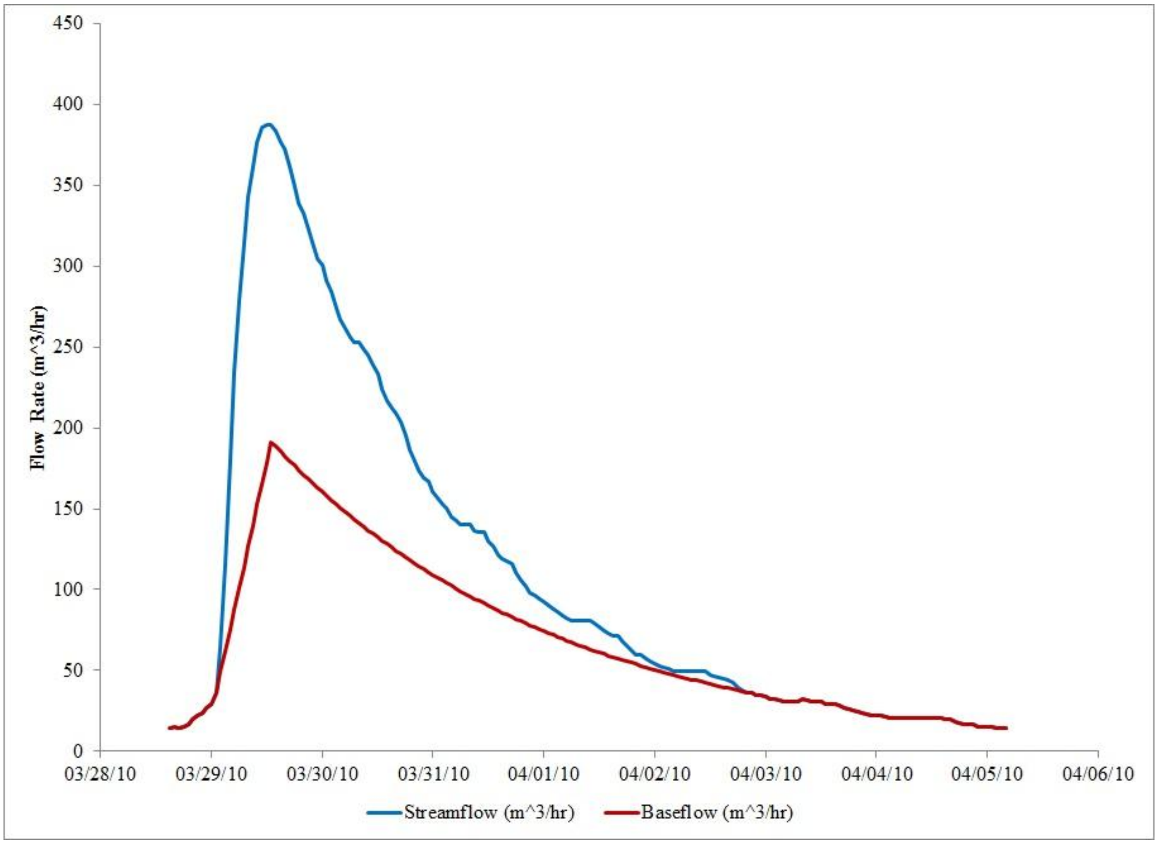

Figure B-17: Hydrograph separation for storm event on 3/28/10.

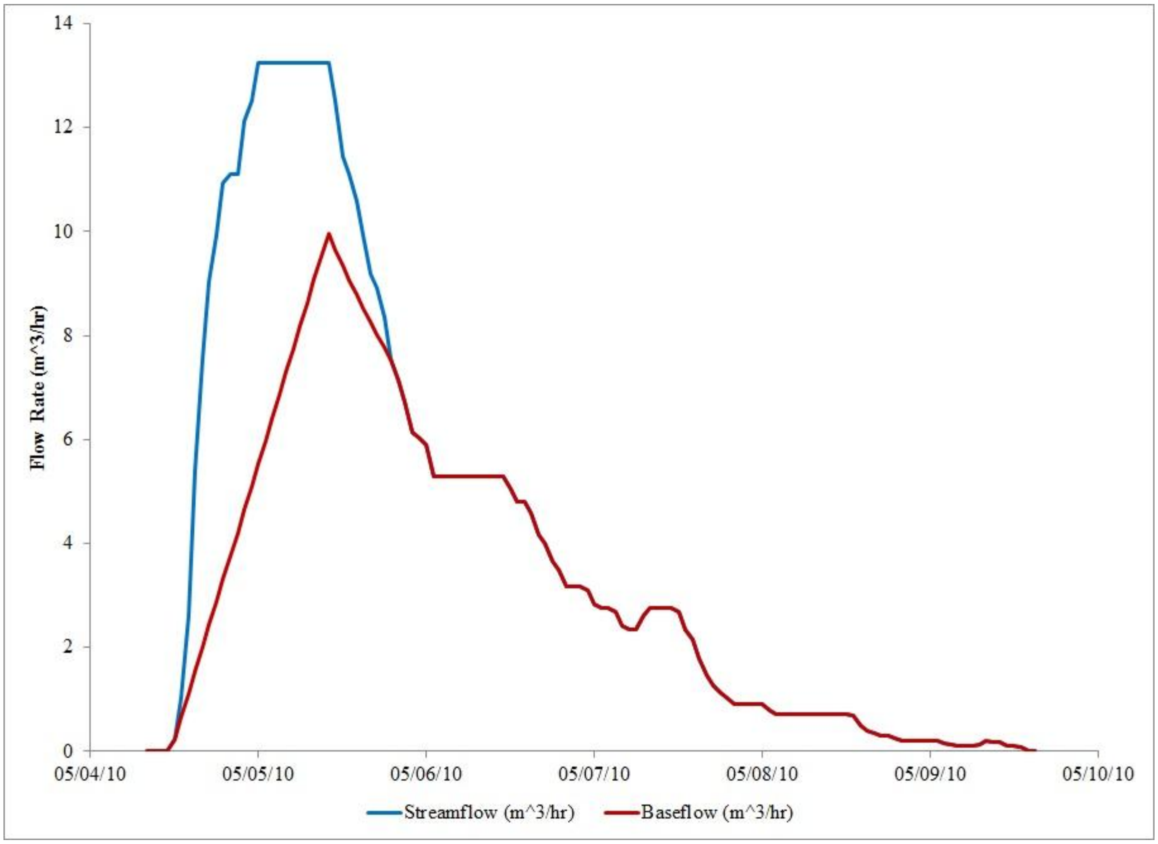

Figure B-18: Hydrograph separation for storm event on 5/4/10.

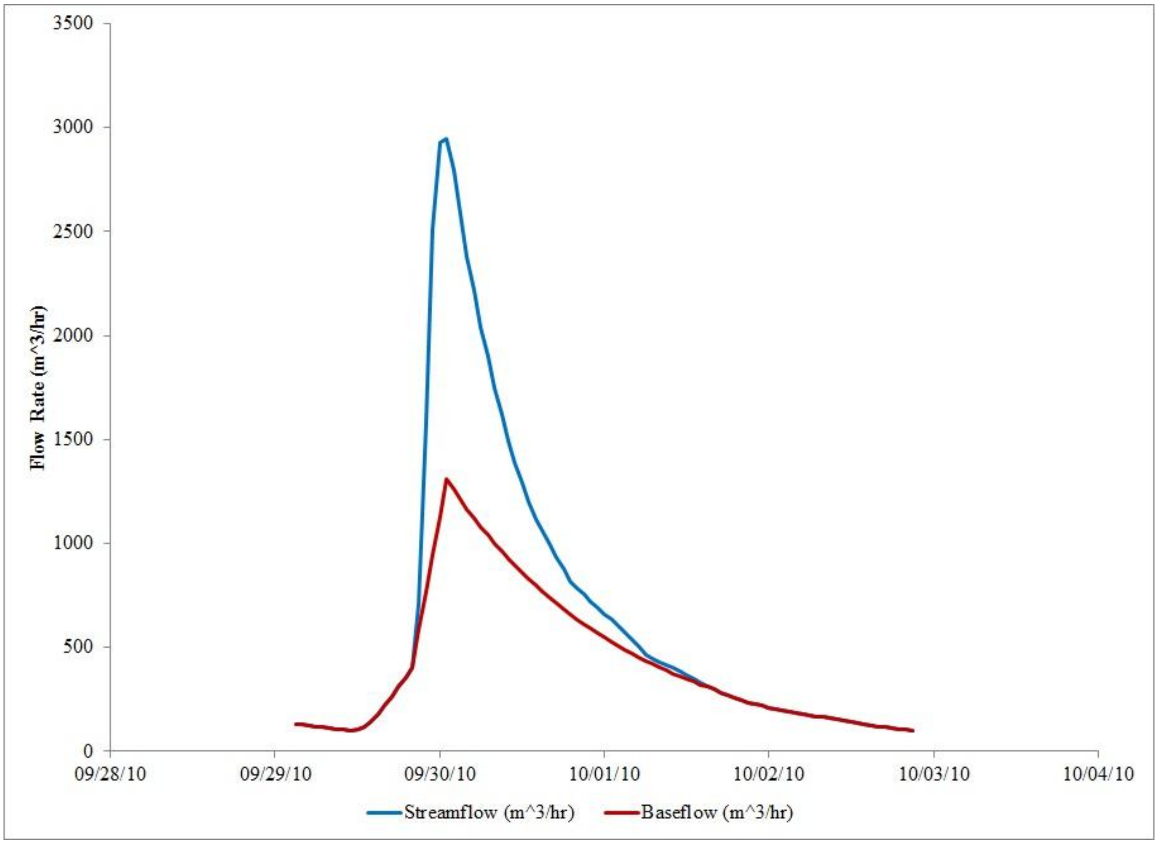

Figure B-19: Hydrograph separation for storm event on 9/29/10.

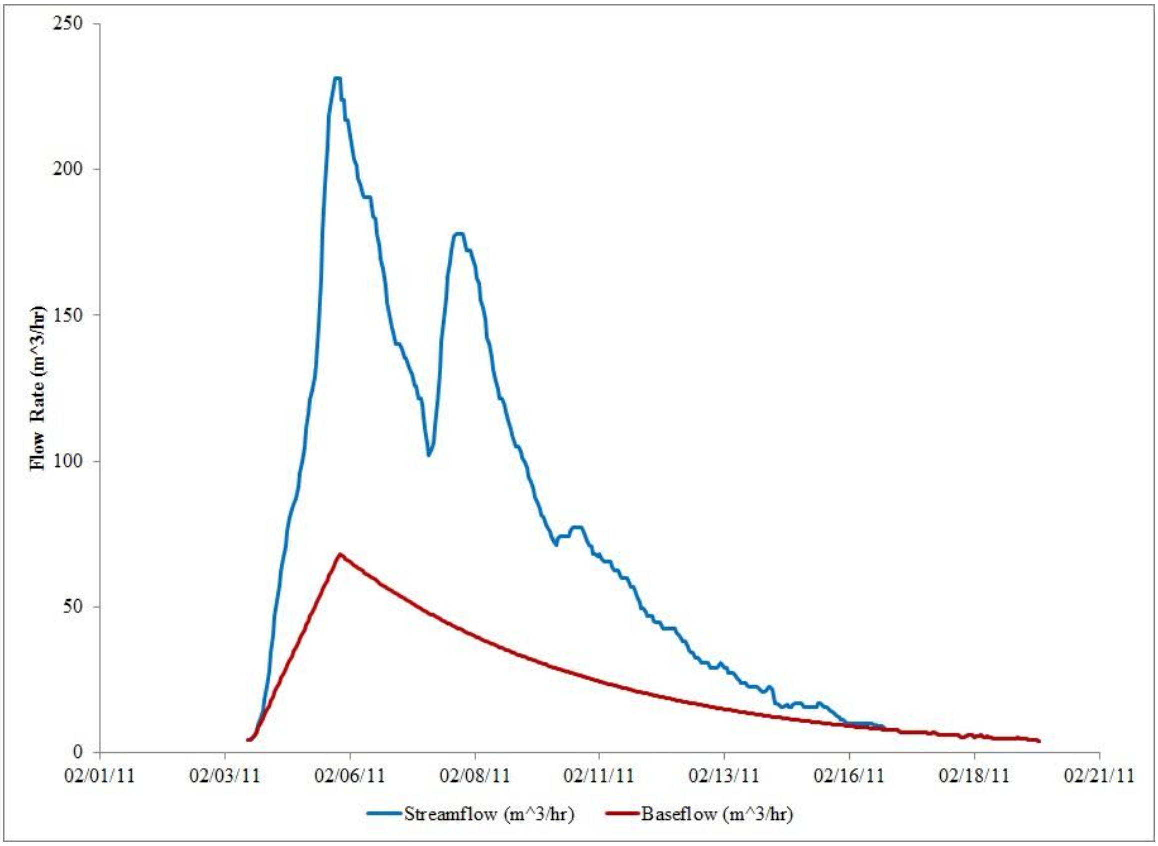

Figure B-20: Hydrograph separation for storm event on 2/2/11.